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**LATE QUATERNARY VEGETATION AND LAKE LEVEL CHANGES
IN CENTRAL ALASKA**

**A
THESIS**

**Presented to the Faculty
of the University of Alaska Fairbanks
in Partial Fulfillment of the Requirements
for the Degree of**

DOCTOR OF PHILOSOPHY

**By
Nancy Horner Bigelow, B.A., M.A.**

**Fairbanks, Alaska
August 1997**

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IN CENTRAL ALASKA


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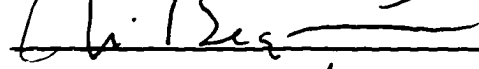
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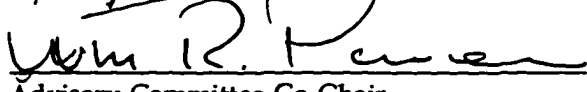


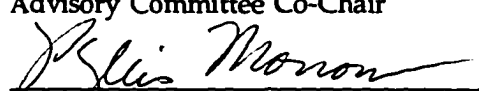






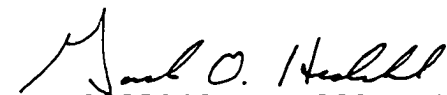


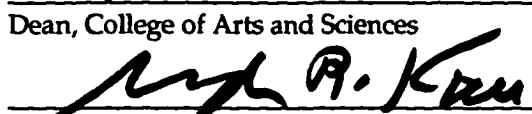
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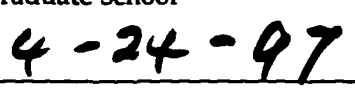
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ABSTRACT

The threat of significant high-latitude global warming over the next 50 years requires that we assess the response of vegetation to climate change. One approach is to see how plants have reacted to past climate change. In this study high-resolution reconstructions of past vegetation and climate, based on pollen and lake level changes, provide useful insights into vegetation and climate change in central Alaska since 14,000 years ago.

Climate changed substantially at about 12,000 years ago, between 11,000 and 10,000 years ago, and about 8000 years ago. At 12,000 years ago, a significant transition is reflected by the appearance of shrub birch into a region that had been dominated by grass, sage, and sedge. The vegetation became denser; shrubs occupied the moister sites, and herbaceous taxa grew on well-drained, exposed ridges and slopes. Lake levels increased at this time, suggesting the climate became warmer and wetter than it had been previously.

Between 11,000 and 10,000 years ago, the vegetation at some sites reverted to a grass and sage-rich flora, suggesting a return to drier and/or cooler conditions. This period of climate change has not been recognized before from pollen records in central Alaska. The timing of this vegetation shift suggests it is related to the Younger Dryas event, a world-wide episode of climatic deterioration.

About 8500 to 8000 years ago, spruce appeared in the region, coincident with a significant lake level rise, suggesting that the spruce expansion was aided by wetter conditions, as well as warmer temperatures.

In central Alaska, periods of past vegetation change are marked by shifts in moisture. Today, central Alaska receives very little rain, and in some areas the vegetation is moisture-limited, suggesting that during the past, changes in moisture could have had a strong effect on the vegetation. In terms of future global change, this study suggests that any shifts in moisture associated with the predicted temperature changes, especially towards drier conditions, will strongly affect the current vegetation distribution.

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ACKNOWLEDGMENTS

This project would not have been possible without the advice and support (both moral and financial) from my advisory committee. Three committee members deserve special mention. Dr. Powers introduced me to the world of Quaternary studies, especially as it applies to archaeological questions. Dr. Edwards taught me the skills to successfully undertake a paleoecological research project, not the least of which being how to core a lake during the winter. Dr. Finney showed me that lakes should not be considered only pollen collectors, and that changes internal to the lake system also provide valuable information. All my committee members were excellent sounding boards for some of my more unconventional ideas (most of which have since landed in the rubbish heap). Any remaining errors or omissions, however, are my own.

Financial support for this dissertation was provided by PALE research grants to Edwards and Finney, as well as grants from the Center for Global Change and the University of Alaska Museum Geist Fund to myself.

Finally, I wish to thank my family, friends, and cats for their patience as I struggled through the last six months while I was writing and tearing my hair out. Fortunately, most of them took it in stride, and ignored me. My cats, however, are another matter.

CHAPTER 1—INTRODUCTION

Introduction

As a result of anthropogenic increases in greenhouse gases, global climate is predicted to change rapidly over the next 50 years (IPCC, 1990). How accurate are the model predictions of climate change? How are high latitudes likely to be affected?

Validating climate simulations is an important aspect of paleoecological research. Accurate paleoecological reconstructions not only provide insights into past climates, but also provide an independent test of the model simulation results, indicating where the models may need adjustment.

At high latitudes and in the boreal forest, climate model simulations suggest that both summer and winter temperatures will increase, as will precipitation, although summers may become more droughty (Weller et al., 1995; Kirschbaum and Fischlin, 1995). How will these changes affect the landscape? Model simulations indicate the immediate effects may be increased forest fire frequency, as well as insect pest outbreaks. The composition of the vegetation will also change, affecting not only herbivore distribution and densities, but also the northern economy because of changes in the distribution of commercially valuable tree species (Armbruster et al., 1995; Criddle, 1995; Kirschbaum and Fischlin, 1995).

Studying how vegetation responded to climate change in the past may shed light on how the vegetation is likely to respond to climate change in the future. Various sorts of proxy climatic records, such as glacial deposits and pollen records indicate that since 18,000 years ago, high latitude climates have been both colder and warmer than today. Accurately describing both the climate and the vegetation during this period will provide useful insights into how high latitudes are likely to be affected by future climate change.

Central Alaska, because of its latitude (approximately 63°N to 66°N) and because of its location (Figure 1.1) between mountain ranges that block moisture from neighboring oceans, is sensitive to small changes in temperature and effective moisture (precipitation-minus-evaporation). Existing paleoecological research suggests temperature and effective moisture

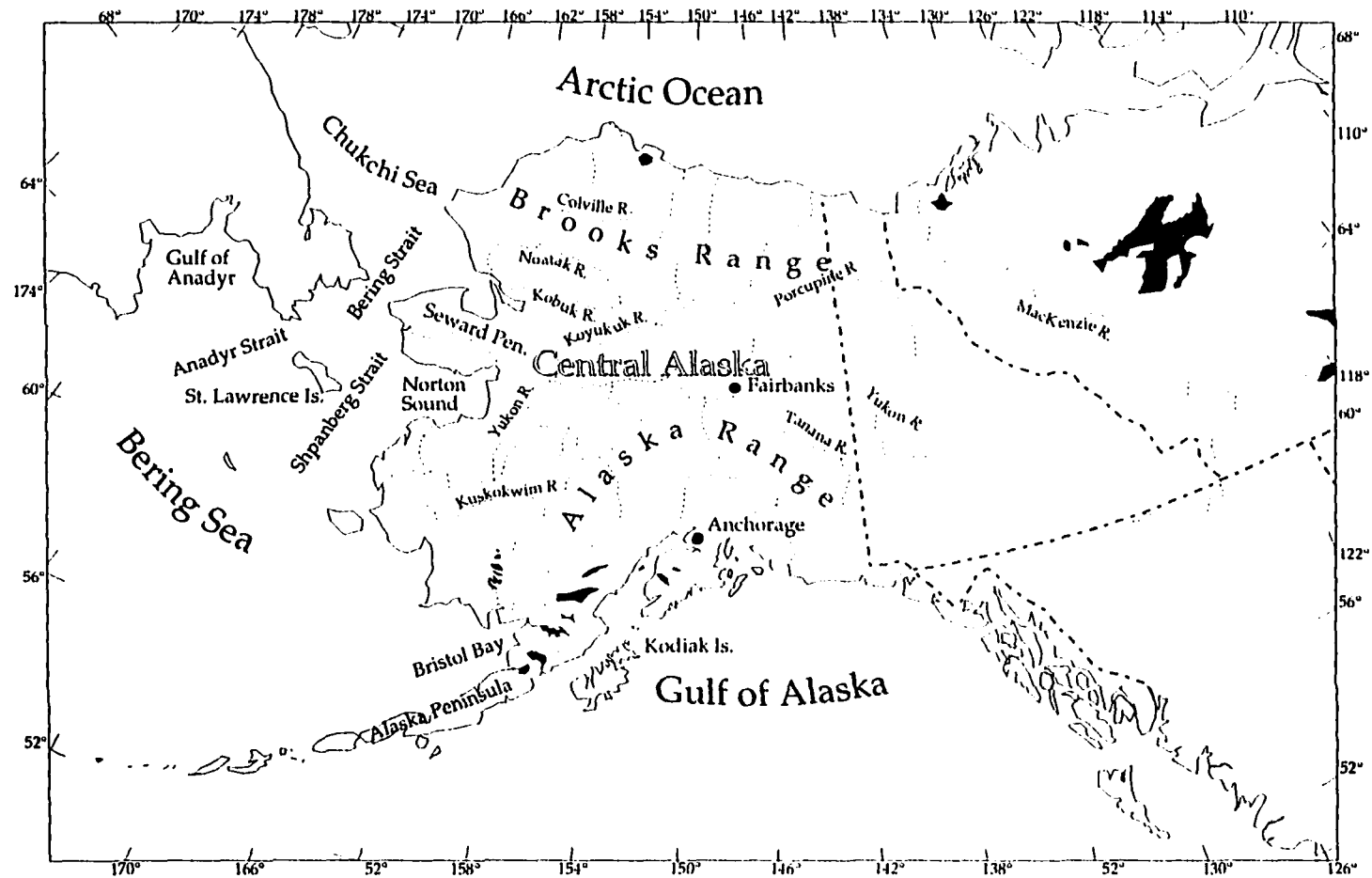


Figure 1.1: Map of locations mentioned in the text.

have changed significantly during the past 14,000 years (Anderson and Brubaker, 1994; Barnosky et al., 1987). Until recently, we have not had high-resolution paleoecological reconstructions capable of reflecting the sort of rapid climate change expected from greenhouse warming. In addition, we have not been able to unravel the separate effects of temperature changes from changes in effective moisture. Finally, there has been no paleoecological research in Alaska that addresses how people would have been affected by climate change. This study addresses these problems by answering the following research objectives.

Research Objectives

- Has central Alaska been affected by short-term (≤ 1000 yr long) climatic perturbations, such as the Younger Dryas?

The Younger Dryas (YD) is a short period (*ca* 1000 yr) of sharply deteriorating climate that occurred in the midst of postglacial climatic amelioration. The YD was originally identified in northwestern Europe (Mangerud et al., 1974). Recently, however, numerous proxy climatic records from around the world suggest this event occurred globally (c.f. Gosse et al., 1995; Clapperton et al., 1997; Gasse et al., 1991; Kudrass et al., 1991).

In Alaska, the YD has been identified primarily in coastal regions (Peteet and Mann, 1994; Engstrom et al., 1990; Hansen and Engstrom, 1996). The current study suggests that non-coastal regions were also affected by this climatic shift. A world-wide YD event is significant because until recently, climate change has been usually interpreted in light of large-scale, slowly changing forcing mechanisms, such as Milankovitch orbital variations (Berger, 1978), or postglacial deglaciation. These forcing mechanisms were then used to create climate model simulations (c.f. CLIMAP, 1976; COHMAP Members, 1988), which were also large-scale with coarse spatial and temporal resolutions. However, we have recently become aware that climate change often occurs over a very short time, at a scale which is beyond the temporal or spatial resolution of the climate simulations and which cannot be adequately

explained by large scale forcing mechanisms such as orbital variations or deglaciation.

The advent of high-resolution sampling of sensitive climate proxies indicate climate can change significantly over the course of a human life time. Ice cores from Greenland indicate the end of the Younger Dryas occurred over 50 years or less (Dansgaard et al., 1989; Alley et al., 1993).

This sort of rapid climate change has not been typically documented in Alaska. Part of this results from the use of records which may respond slowly to climate change (such as glaciers). In addition, many records have a very coarse chronological resolution, making it impossible to identify climatic changes that are ≤ 1000 years in duration.

- Is there evidence of changes in effective moisture ?

Organisms respond to the moisture that is available to them. For this reason, effective moisture (precipitation-minus-evaporation) is the variable used in this study. While precipitation obviously affects the amount of moisture on the landscape, temperature changes can be a significant mitigating factor, resulting in dry episodes during periods of high precipitation, or the reverse.

All the proxy climatic data from central Alaska indicate the full glacial was cold and dry, while the Holocene was warmer and moister (c.f. Guthrie, 1990a; Ritchie, 1982; Anderson and Brubaker, 1994). We do not know whether temperature and effective moisture changed simultaneously. If the climate warmed before effective moisture increased, then there may have been a period when central Alaska was significantly drier than today. In addition, we do not know whether the documented changes in postglacial vegetation are due to increasing warmth, or effective moisture, or both.

Distinguishing changes in effective moisture from changes in temperature is difficult, because most climatic proxies are sensitive to both variables. Lake levels address this problem because they respond to the regional moisture budget. That is, they represent an equilibrium between rainfall on the lake, catchment run-off, groundwater flux, and evaporation from the lake surface (Harrison and Digerfeldt, 1993). However, because effective moisture is itself

affected by temperature, lake level changes actually integrate precipitation and temperature together. It is only when lake level records can be combined with a temperature record (i.e., Chironomid head capsules [c.f. Walker, et al., 1991]) that temperature and precipitation can be untangled.

- How might have people been affected by climate change?

Climate change affects humans primarily through changes in landscape productivity, although the human response can be highly varied, ranging from migration, to changes in subsistence strategies, and/or changes in social organization (McGovern, 1991; Mason and Gerlach, 1995; Powers and Jordan, 1990).

Whether climate change resulted in cultural transitions and/or economic reorganization, most studies of human/climate interaction can be summarized by:

"it got cold/hot/dry/wet and they
died/fluoresced/migrated/intensified" (McGovern, 1991:78).

One important problem is the climate data are usually at too coarse a scale (both temporally and geographically) to be of much use to the archaeologist. That is, people choose to live at a site because of the conditions *there* and *at that time*, not the conditions 50 km away and 500 years later. In addition, the number of archaeological sites on the landscape for any given time period tends to be low. As a result, archaeologists have difficulty explaining the variation between sites, whether is due to changes in site use, or due to different groups occupying the sites.

The specific linkages between climate change and humans can only be reconstructed in regions that have abundant archaeological remains and accurate, detailed climate records (McGovern, 1991; Mason and Gerlach, 1995). In practice, this sort of analysis is limited to the last several thousand years of the archaeological record where historical and high-resolution climatic records can be combined.

This study attempts to reduce some of the ambiguity in the relationship between people and climate by studying climate records at a high temporal

resolution, in addition to collecting the records from areas close to archaeological sites. However, because of the uncertainties inherent in the climatic reconstructions, in addition to the low number of archaeological sites, especially those with faunal remains, this reports makes no attempt to state definitively the linkages between people and climate. Instead, I suggest only how the documented climate changes could have affected landscape productivity, resource distribution/abundance, and hence, people.

Organization of the Dissertation

This dissertation is a hybrid between a series of several publishable papers and a single monograph. To reduce repetition between chapters, methods common to several sites were placed in a separate appendix (Appendix A). Methods that were unique to a particular site (i.e., the modern pollen from Bonanza Bluff) are included in the section that discusses that site.

The chapters on Windmill, Dune, and Birch lakes are essentially data chapters, although they have extensive discussions. Some topics are important at more than one site. In this case, the topic is discussed most fully in one of the data chapters and is summarized in the other data chapters. In addition, one topic (the Younger Dryas) has its own chapter, because data from two of the lakes, as well as the modern pollen analysis contribute to this topic.

The final chapter compares the data among the three lakes and discusses a regional climate forcing mechanism. Afterwards, I summarize the human prehistory of central Alaska, and subsequently outline how climate change would have affected the human populations.

Finally, general background information about central Alaska today, as well as climate and landscape changes since 16,000 ^{14}C yr BP are presented below.

Regional Setting

As a state, Alaska spans a wide diversity of topography and climate. The focus of this study is central Alaska in general, and more specifically, eastern central Alaska. Central Alaska is the region between the Alaska Range in the south, the Brooks Range in the north, the Canadian border in the east and the

Seward Peninsula and the Yukon-Kuskokwim delta in the west (Figure 1.1). Eastern central Alaska is the eastern part of the region, from the Canadian border to about 152° W longitude (Figure 1.1).

Climate. The mean annual temperature at Fairbanks, located in eastern central Alaska, is -3.5°C, with a mean winter (DJF) temperature of -22.4°C and a mean summer (JJA) temperature of 14.7°C. The mean annual precipitation is 280.5 mm, with most of it falling as rain during the late summer and early autumn (based on a 30-yr record cited in Rieger et al., 1979). At lower latitudes, the low precipitation and warm summers would make Fairbanks a desert. However, because of widespread permafrost, the surface of which melts during the summer, much of the low elevation areas are in fact poorly drained. In contrast, hillsides, especially those facing south, can be moisture-limited because of the high solar radiation and low precipitation (Wesser and Armbruster, 1991; Lloyd, 1993).

Topography. The topography of interior Alaska is a mixture of flat alluvial plains with high mountains and rolling hills. The alluvial plains, particularly the Tanana and Yukon basins, are underlain primarily by fine-grained silt and sand. These thick deposits, particularly those in the lowlands, contain abundant ice-rich permafrost, which impedes drainage, and causes widespread paludification.

The hills are mantled with wind-deposited loess. The south-facing hill sides receive abundant solar radiation and are not underlain by ice. In contrast, the north-facing slopes are much colder and can contain significant amounts of ice. At higher elevations, the loess mantle is thin or absent, as is ice-rich permafrost.

Vegetation. The vegetation of interior Alaska is strongly affected by topography, successional stage, and forest fire history (Hultén, 1968; Viereck 1970, 1983; Viereck et al., 1992).

The hillsides are vegetated with pure or mixed stands of aspen (*Populus tremuloides*), birch (*Betula papyrifera*), white spruce (*Picea glauca*), and black spruce (*Picea mariana*). Microsite temperature and moisture variations, in addition to successional history strongly affect species distributions. South-facing slopes are dominated by *P. tremuloides*, *B. papyrifera* and *P. glauca*. In these areas, fire history plays an important role, and localities which are

burned repeatedly may remain *P. tremuloides* and *B. papyrifera*-dominated (Quirk and Sykes, 1971), while areas more rarely burned eventually succeed to a *P. glauca* forest.

North-facing slopes, because of inadequate drainage and shallow permafrost, are dominated by scattered *P. mariana*, larch (*Larix laricina*), shrub birch (*Betula glandulosa*) and *Salix* spp. Trees are rare or absent in areas with very poor drainage, and the vegetation is dominated by *Carex* spp., cotton grass (*Eriophorum* spp.), labrador tea (*Ledum groenlandicum*), and cassandra (*Chaemadaphne calyculata*).

Steep south-facing slopes are droughty localities and are often treeless or have only a scattering of *P. tremuloides*. The vegetation is dominated by grasses, sedges, and forbs such as *Aster* spp. (daisies) and *Pulsatilla patens* (pasque flower), along with low-lying shrubs, such as *Artemisia frigida* and *Rosa acicularis*. Juniper bushes (*Juniperus communis*) are also present in some localities.

Depending on successional stage and forest fire frequency, the vegetation on the alluvial plains is a mosaic of *Salix* spp., *Alnus*, balsam poplar (*Populus balsamifera*), *P. glauca*, *P. mariana*, and larch (*Larix laricina*). Early successional communities are dominated by *Salix*, *Alnus*, and *P. balsamifera*. As the *P. balsamifera* dies and leaves openings in the canopy, *P. glauca* begins to dominate. On older alluvial deposits, the *P. glauca* forest develops into a *P. mariana* forest, and finally into a bog (Viereck, 1970; Viereck et al., 1986). The hypothesized mechanism forcing this final transition is a thickening of the organic mat, resulting in cooler soil temperatures, higher permafrost table, and impeded drainage, which promotes the development of a *P. mariana* bog. This successional sequence can be strongly affected, however, by repeated burning, as well as flooding, so that areas with a high flood frequency may remain a *P. glauca* forest indefinitely (Mann et al., 1995).

Alaska 16,000 ¹⁴C yr BP to the Present

Climate model simulations and paleoecological research in central Alaska indicate climate has changed significantly since the last glaciation. The section below summarizes climate model simulations and various proxy climatic records for the region. It is difficult to compare these two types of

data because of differences in their chronologies. Model simulations are produced in calendar years before the present (cal yr BP). In contrast, most proxy climatic data sets are in *radiocarbon* years before 1950 A.D. (^{14}C yr BP).

Radiocarbon years are based on the quantity of ^{14}C in organic materials of various ages. The amount of ^{14}C in the atmosphere, however, has not been constant (Taylor, 1987). As a result, radiocarbon years diverge significantly from calendar years, so that by 14,000 cal yr BP, the comparable radiocarbon age is about 12,000 ^{14}C yr BP (Stuiver and Reimer, 1993)). One way to alleviate this problem is to calibrate the radiocarbon time-scale by radiocarbon dating tree rings of known calendar age. For time periods beyond the limit of the tree ring chronology, radiocarbon dates on corals of known uranium-thorium age also provide a calibration (Stuiver and Reimer, 1993). These transformations yield an age that is thought to be similar to calendar years, although strictly speaking, the calibrations in fact yield ages in tree ring years and in uranium-thorium years. For simplicity's sake, I have not distinguished between calendar years and calibrated years, and both are abbreviated as cal yr BP.

Climate model simulations. The climate model simulations discussed below result from calculations that use solar radiation, as well as global-scale boundary conditions (i.e., ice extent, sea surface temperature, and atmospheric CO_2) to retrodict large-scale atmospheric circulation (Kutzbach et al., 1993; COHMAP Members, 1989). However, changes in solar radiation are measured in cal yr BP, while the boundary conditions are in ^{14}C yr BP. As a result, the model simulations reflect a mixture of the two chronological scales. Thus, discussions of insolation changes are in calendar years, while paleoclimatological or paleoecological changes are in radiocarbon years.

At 18,000 ^{14}C yr BP, an extensive Laurentide ice sheet diverted most of the eastward-flowing jet stream south of the ice sheet and created a strong blocking anticyclone over central Canada. This resulted in increased southerly flow into Alaska and a strengthened Aleutian Low. Summer and winter temperatures were significantly colder at the ice sheet edge, although the simulation suggests that much of central Alaska may not have been much cooler than today. (Glaciological and paleoecological studies however, indicate substantial cooling during this period [Hamilton, 1986b; Ritchie, 1982;

Cwynar, 1982]). Winter and summer effective moisture, were significantly less, possibly about 1/2 modern values (COHMAP Members, 1988; Kutzbach et al., 1993).

By about 12,000 ^{14}C yr BP, climate model simulations indicate a weakened Laurentide anticyclone, with a diverted/split jet stream only in summer. These features would reduce the southerly flow into Alaska, and increase the westerly flow from Asia. In central Alaska, temperature (winter and summer) and effective moisture were not significantly different from modern levels, except for the eastern part of the region, where effective moisture and temperature were still significantly reduced (COHMAP Members, 1988; Kutzbach and Ruddiman, 1993; Kutzbach et al., 1993).

From 15,000 to 8000 cal yr BP (13,000 to 7000 ^{14}C yr BP), orbital variations resulted in increased seasonality at 65°N , reaching maximum values between about 12,000 and 9000 cal yr BP (11,000 to 8000 ^{14}C yr BP). At this time, summer insolation was approximately 9% greater than present, and winter insolation was about 25% less than present (Berger, 1978; Bartlein et al., 1991). Climate model simulations at 9000 cal yr BP (8000 ^{14}C yr BP) suggest that summer temperatures were significantly higher than the present, although winter temperatures were only slightly (and insignificantly) cooler. Effective moisture was not significantly different from today (Kutzbach et al., 1993). By this time, the Laurentide ice sheet had shrunk considerably, and only had minor temperature or circulation effects on central Alaska (Bartlein et al., 1991).

At about 6000 cal yr BP (6000 ^{14}C yr BP), summer insolation at 65°N was still greater than modern, resulting in significantly warmer simulated temperatures (Berger, 1978; Kutzbach et al., 1993). After 6000 cal yr BP, orbital effects relaxed, and regional climatic conditions were probably close to modern by about 4000 yr BP (Bartlein et al., 1991).

Glacial history. During the last glacial maximum, about 18,000 ^{14}C yr BP, central Alaska was predominantly unglaciated because the region was too arid for glacial growth. Instead, glaciers were limited to the mountain valleys of the Brooks Range and Alaska Range, and did not extend beyond the foothills (Hamilton 1986a, 1986b; Péwé, 1975; Wahrhaftig, 1958; Ten Brink, 1983).

By about 16,000 ^{14}C yr BP, ice was retreating in all mountain systems and reaching near-Holocene volumes by about 14,000 ^{14}C yr BP. Subsequently, brief glacial advances in the Brooks Range and Alaska Range date to roughly 13,500 to 14,500 ^{14}C yr BP (Hamilton, 1986b). Two subsequent and small advances occurred in the Alaska Range about 12,500 ^{14}C yr BP and sometime between 9500 and 10,400 ^{14}C yr BP (Ten Brink and Waythomas, 1985).

After about 4000 ^{14}C yr BP, high elevation cirque glaciers readvanced in the Brooks and Alaska ranges, probably in response to minor climate fluctuations (Calkin, 1988; Hamilton, 1986b).

Vegetation history. Before 14,000 ^{14}C yr BP, pollen records (Figure 1.2) suggest the landscape was sparsely vegetated with Poaceae, Cyperaceae, and *Artemisia* (Anderson and Brubaker, 1994). The exact nature of this landscape is controversial, mainly because there is no close modern analogue for a vegetation with high abundances of these taxa, especially *Artemisia*. One school of thought (e.g. Guthrie, 1990a) views this grassland as highly productive, with similarities to temperate steppes. Faunal remains from large herbivores (mainly grazers) are abundant during this period, indicating that a productive grassland must have existed to support the animal populations.

Another school of thought (e.g. Ritchie, 1982, Cwynar, 1982) suggests this vegetation was a barren unproductive Arctic tundra. Comparisons between fossil pollen from sites in Alaska and Canada with modern collections from a variety of settings indicates the high Arctic islands of Canada are the most similar to the ancient pollen assemblages (Anderson et al., 1989).

Between 14,000 ^{14}C yr and 12,000 ^{14}C yr BP, the vegetation changed dramatically with the appearance of *Betula* (presumably shrubby species [*B. glandulosa* or *B. nana*]). The exact timing of this transition is problematic. At Birch Lake (discussed in Chapter 4), where the transition has been dated by AMS and conventional methods, the AMS age is about 2000 ^{14}C yr younger than the original estimate. Pollen influx studies suggest that while *Betula* became an important component of the vegetation, most herbaceous taxa from the previous period remained abundant, indicating that the new vegetation was probably a mosaic, with *Betula* occupying the mesic sites, and

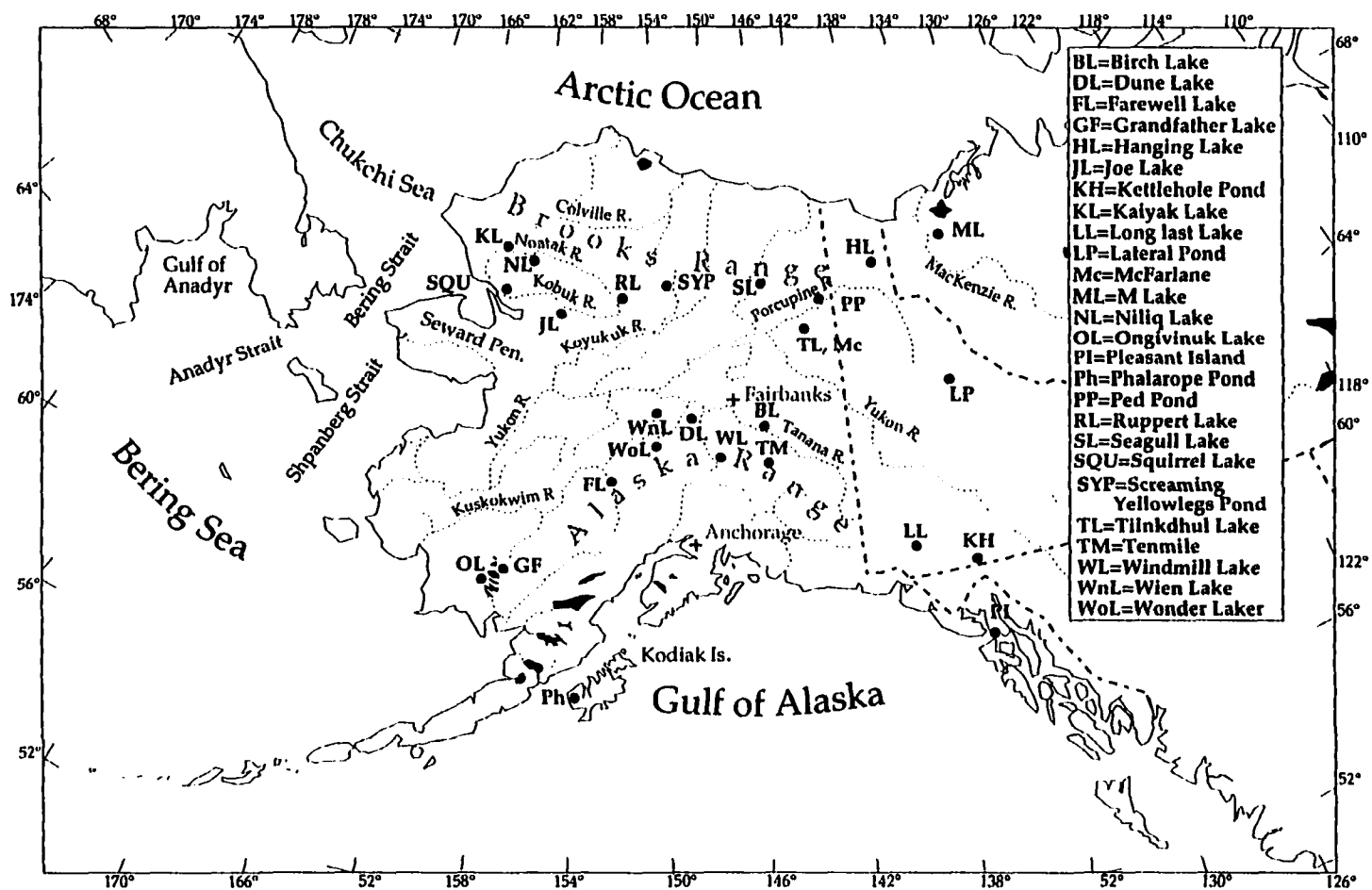


Figure 1.2: Location of pollen records mentioned in the text.

herbaceous taxa occupying the more arid and exposed sites (Ritchie, 1982; Anderson and Brubaker, 1994).

From roughly 12,000 to 9000 ^{14}C yr BP, many pollen records indicate that *Populus* became locally abundant on the landscape (Edwards and Barker, 1994; Bartlein et al., 1995). The exact nature of the vegetation is unclear, partly because we are not certain whether the pollen represents *P. balsamifera* or *P. tremuloides*. In most cases, *P. balsamifera* is the most likely candidate because in a few sites where the pollen has been identified to the species level (Brubaker et al., 1983; Cwynar, 1988), *P. balsamifera* is the most common type. In addition, *P. balsamifera* leaves dated to almost 14,000 ^{14}C yr BP have been recovered from northern Alaska (Hopkins et al., 1981). It is probable that at most sites the *Populus* pollen represents an expansion of gallery forests along rivers and around lakes (Anderson and Brubaker, 1994; Anderson et al., 1988). Some cores also show an expansion of *Salix* at this time, suggesting both species were common in such settings, as they are today.

About 9000 ^{14}C yr BP, *Picea* appeared in the region. There is clear evidence that this taxon migrated rapidly from southern Canada to Alaska after the destruction of the Laurentide ice sheet (Ritchie and MacDonald, 1986). This is also supported by its diachronous appearance across Alaska, with the earliest sites in the east. *P. glauca* was probably the initial migrant, possibly colonizing well-drained sites where *Betula* may not have been abundant.

About 7000 to 6000 ^{14}C yr BP, *P. mariana* and *Alnus* become common on the landscape in central Alaska. Rare pollen grains suggest that *Alnus* was probably always present earlier in the Holocene, but at low densities and in scattered locations. By 7000 ^{14}C yr BP, however, it rapidly expanded from the western Brooks Range to the Canadian border (Anderson and Brubaker, 1994).

After 6000 ^{14}C yr BP, all the modern constituents of the boreal forest were present in the region, and with minor variations in relative abundances, the vegetation was broadly similar to modern, at least after 4000 ^{14}C yr BP (Anderson and Brubaker, 1994).

Human prehistory. The earliest well-dated archaeological sites in Alaska date from about 12,000 to 11,500 ^{14}C yr BP (Hoffecker, et al., 1993). The sites, (all of which are apparently hunting overlook sites), extend from the Brooks

Range (Mesa) to the Alaska Range (Dry Creek, Walker Road, Broken Mammoth, Swan Point, and others [Figure 1.3]). Faunal remains are extremely scarce, but the bones recovered from Broken Mammoth indicate the people were relying on migratory birds, bison, and elk (Kunz and Reanier, 1994; Powers and Hoffecker, 1989; Yesner et al., 1992; Yesner, 1996; Holmes et al., 1996; Cook, 1996).

By roughly 10,500 to 8000 ^{14}C yr BP, changes in the lithic technology suggest either a different group of people, or a change in hunting strategies in central Alaska (Powers and Hoffecker, 1989; West, 1981). Faunal remains, while still scarce, suggest that the people were hunting mainly bison, elk, and caribou (Yesner et al., 1992; Yesner, 1996; West, 1981).

From 8000 to 6000 ^{14}C yr BP, archaeological sites become quite scarce, with no sites in central Alaska that date between 7000 and 6000 ^{14}C yr BP. It is unclear whether this reflects reduced human populations (West, 1981, 1996) or reduced preservation/visibility of their activities.

After 6000 ^{14}C yr BP, sites are found in a variety of locations, in the uplands, along lakes and rivers, and at overlooks. Faunal remains indicate caribou was the important prey in the earlier part of the period, although by the end of the Holocene, fish and moose were also significant resources (Betts, 1987; Sheppard et al., 1991; Holmes, 1986; Dixon, 1985).

Summary

Climate is expected to change significantly over the next 50 years. Paleoecological studies of past climates help our understanding of future climate change by providing independent tests of climate model simulations. The boreal forest and high latitudes are expected to be severely affected by global warming. Currently, the paleoecology of the region is too temporally coarse to adequately test model simulations for the 50 yr time scale expected from global warming. In addition effective moisture (which may change significantly due to global warming) has not been previously reconstructed.

Central Alaska was chosen as the study area because the region is sensitive to small changes in temperature and effective moisture. Today, central Alaska has a varied topography of wide alluvial plains interspersed with

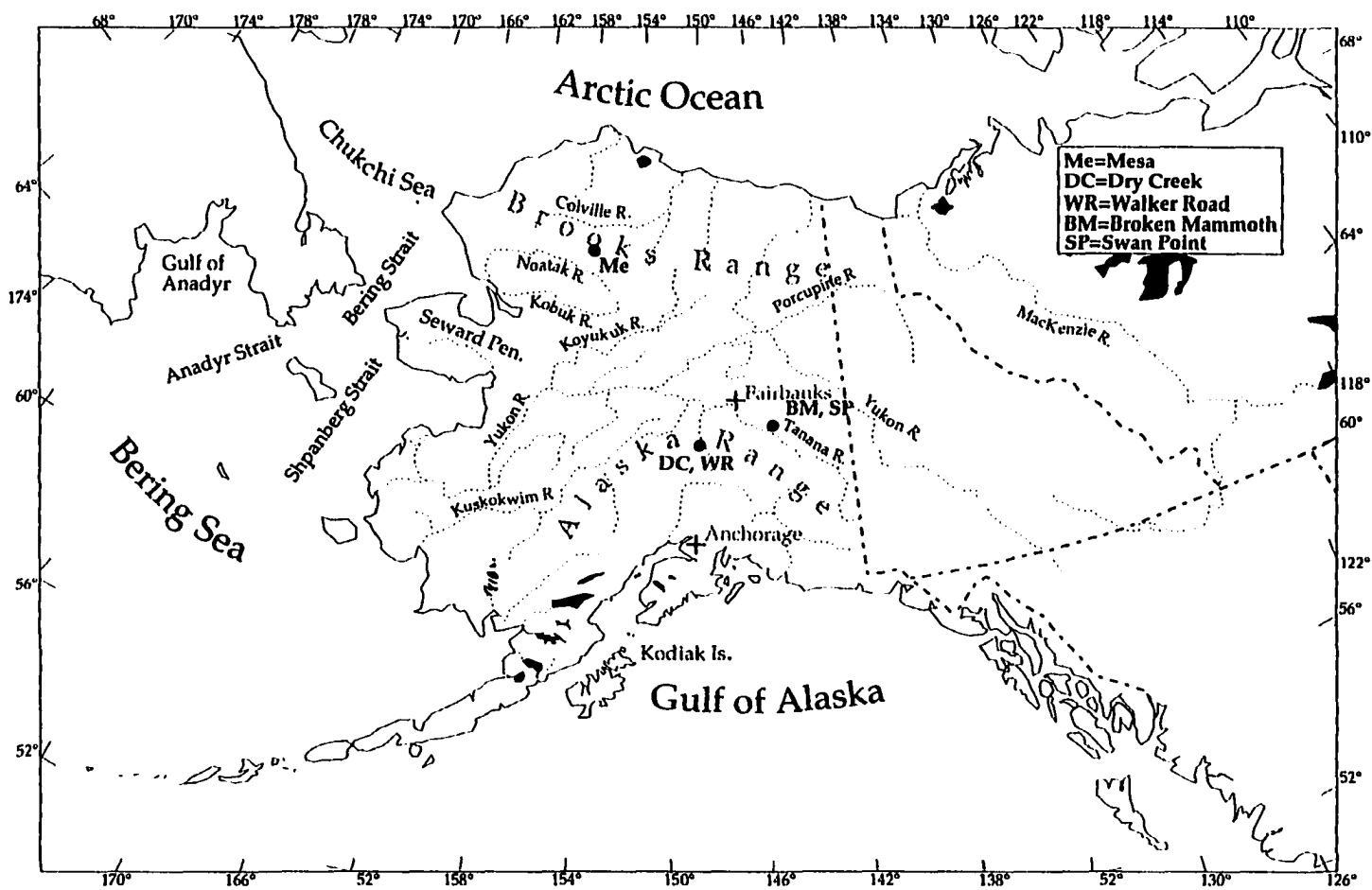


Figure 1.3: Map of archaeological sites mentioned in the text.

rolling hills and high mountains. The vegetation varies according to topography, forest fire history, and successional stage.

The region was largely deglaciated during the last ice-age. The vegetation since that time was initially a sparse herb tundra, which transformed into a shrub tundra about 12,000 ^{14}C yr BP. Trees did not become common until about 8000 ^{14}C yr BP, and the modern vegetation probably did not develop until after about 4000 ^{14}C yr BP.

People initially occupied the region about 12,000 ^{14}C yr BP. Important prey species were bison and elk. By about 6000 ^{14}C yr BP, caribou became more important, while the bison and elk gradually diminished. During the late Holocene, caribou, moose, and fish were the main resources.

CHAPTER 2—WINDMILL LAKE

Introduction

A record from the Nenana valley (Figure 2.1) is critical to this study. The Nenana valley, because it transects the Alaska Range, is more strongly influenced by weather systems in the Gulf of Alaska than other regions in central Alaska. Modern precipitation in central Alaska is strongly influenced by sea surface temperatures and by the strength and position of the Aleutian Low in the North Pacific ocean (Bowling, 1979). As a result, a paleoecological record from the Nenana valley is ideally suited to record climatic shifts that may be the result of changes in North Pacific circulation.

The Nenana valley is also noteworthy for the number of archaeological sites found there. Numerous prehistoric sites have been discovered, five of which are older than 8000 ^{14}C years (Powers and Hoffecker, 1989; Bowers, 1980), indicating the valley has been an important hunting locality for a very long time. A detailed reconstruction of the paleoecology of the valley addresses questions such as why people chose to live there and how the landscape would have appeared to its earliest human inhabitants.

Study Site

Windmill Lake (63°39'N, 148°48'W) is located in the upper reaches of the Nenana valley, approximately 20 km south of the town of Healy (Figure 2.1). Its elevation is about 640 m; the local spruce tree-line is about 800 m above sea level. The lake lies in a diabase bedrock basin with a probable morainal dam at its eastern end.

The region was extensively glaciated during the Pleistocene; ice from the late Wisconsinan glaciation (locally named the Riley creek glaciation) overran the Windmill Lake basin, and reached its maximum extent about 18,000 ^{14}C yr BP approximately 15 km down valley (Wahrhaftig, 1958; Ten Brink, 1983; Ten Brink and Waythomas, 1985). The basal radiocarbon date from Windmill Lake suggests that deglaciation occurred before 12,000 ^{14}C yr BP.

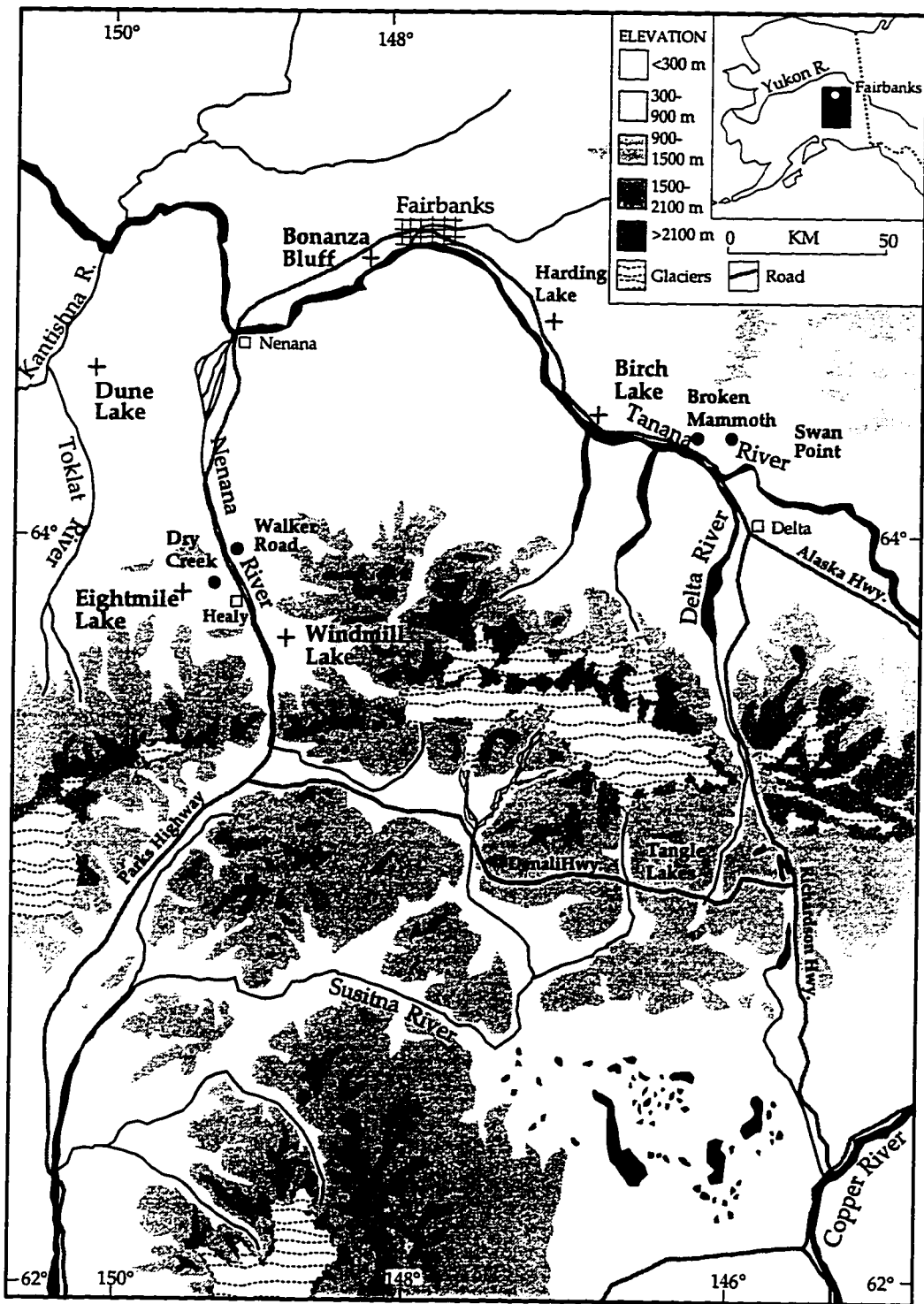


Figure 2.1: Map of locations in the study area.

The local vegetation is mixed *Betula* (trees and shrubs), *Picea* (mainly *P. glauca*), rare *Alnus*, and various *Salix* species. Poorly drained areas also have various ericads (*Ledum* spp., *Vaccinium* spp.) and sedge tussocks. A sedge mat (mainly *Carex* c.f. *rostrata*) surrounds the lake, and aquatic macrophytes include *Nuphar polysepalum* (waterlily), *Potamogeton* c.f. *richardsonii*, and *Chara*. Aquatic mosses (c.f. *Drepanocladus* spp., c.f. *Fontinalis* spp., and others) are abundant to about 75 cm water depth.

The lake is a single basin, approximately 200 m long and 100 m wide, with a maximum depth of about 4 m (Figure 2.2). There is no inlet, although a small seepage and narrow (10 cm wide) creek drain the lake at its southeastern end. During the summer, (July, 1994) the secci depth was about 2.35 m and pH was about 8.0. The surface water temperature was about 18 °C. Temperature and dissolved oxygen profiles taken in July, 1996 indicate the lake is weakly stratified, with the thermocline at about 3 m depth. Dissolved oxygen is low at 3.5 m water depth (2.78-4.34 mg/l), although the lake is apparently not anoxic. Windmill Lake is nestled in its basin, with hills extending between 30 m and 100 m above the lake surface on its south, west, and north sides; the lake is open to the northeast. With the exception of a gap in the hills on the southeast side, the lake is mostly protected from strong winds.

Results

Three cores were retrieved from Windmill Lake: a central core (Windmill B, 410 cm water depth) and two side cores (Windmill C and Windmill D, 158 cm and 85 cm water depth).

Core sediment descriptions. The central core (Windmill B) and the two side cores (Windmill C and D) are 460 cm, 329 cm and 340 cm long, respectively.

Organic-rich gyttja dominates Windmill B from the core top to about 700 cm (all depths are from the water surface) (Figure 2.3). Below about 700 cm, the core gradually becomes siltier, until the core base, which is dominated by sand and pebbles. A tephra layer from the Mount Hayes vent (Jarvis Creek

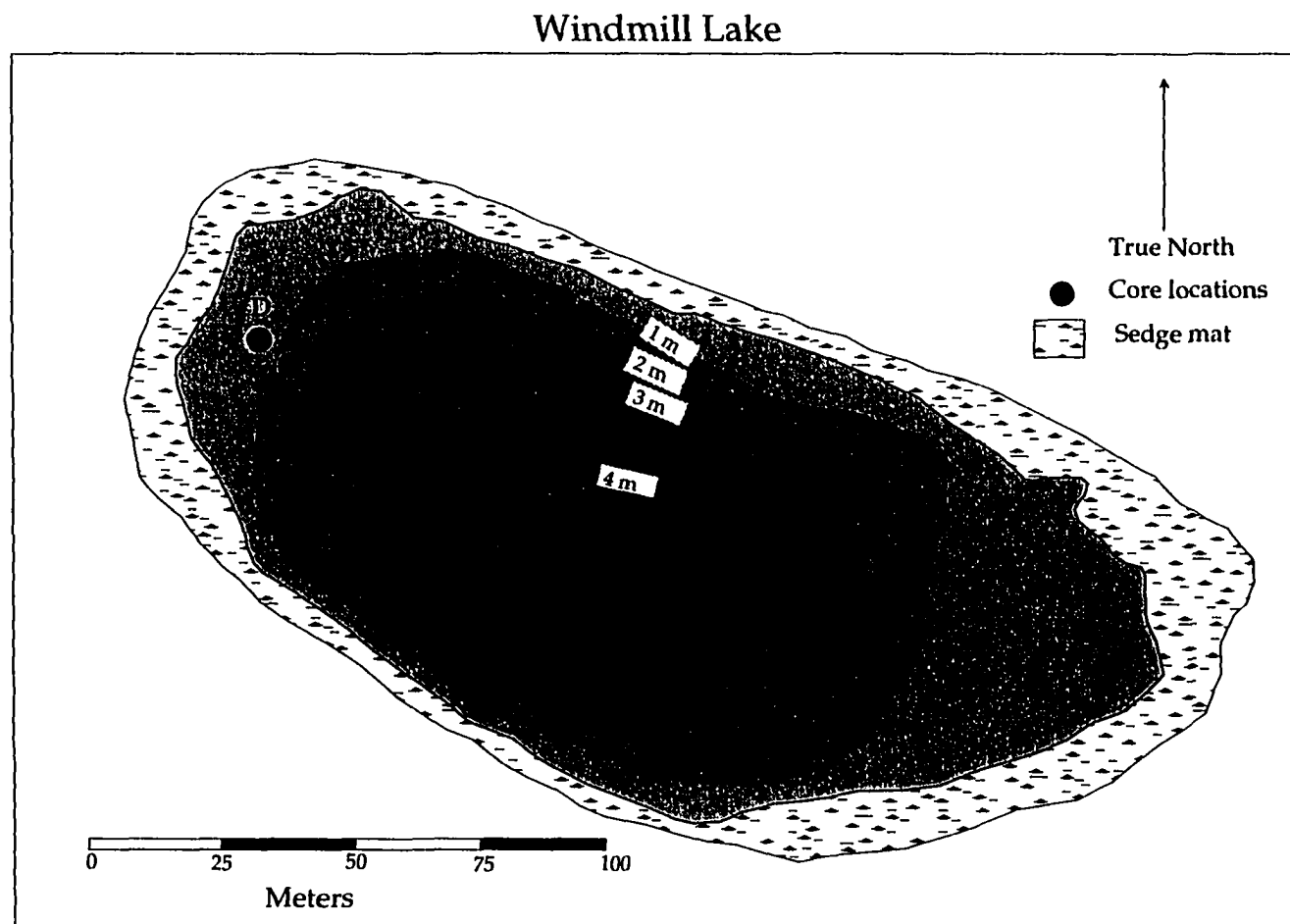


Figure 2.2: Windmill Lake bathymetry and core locations.

Windmill Lake

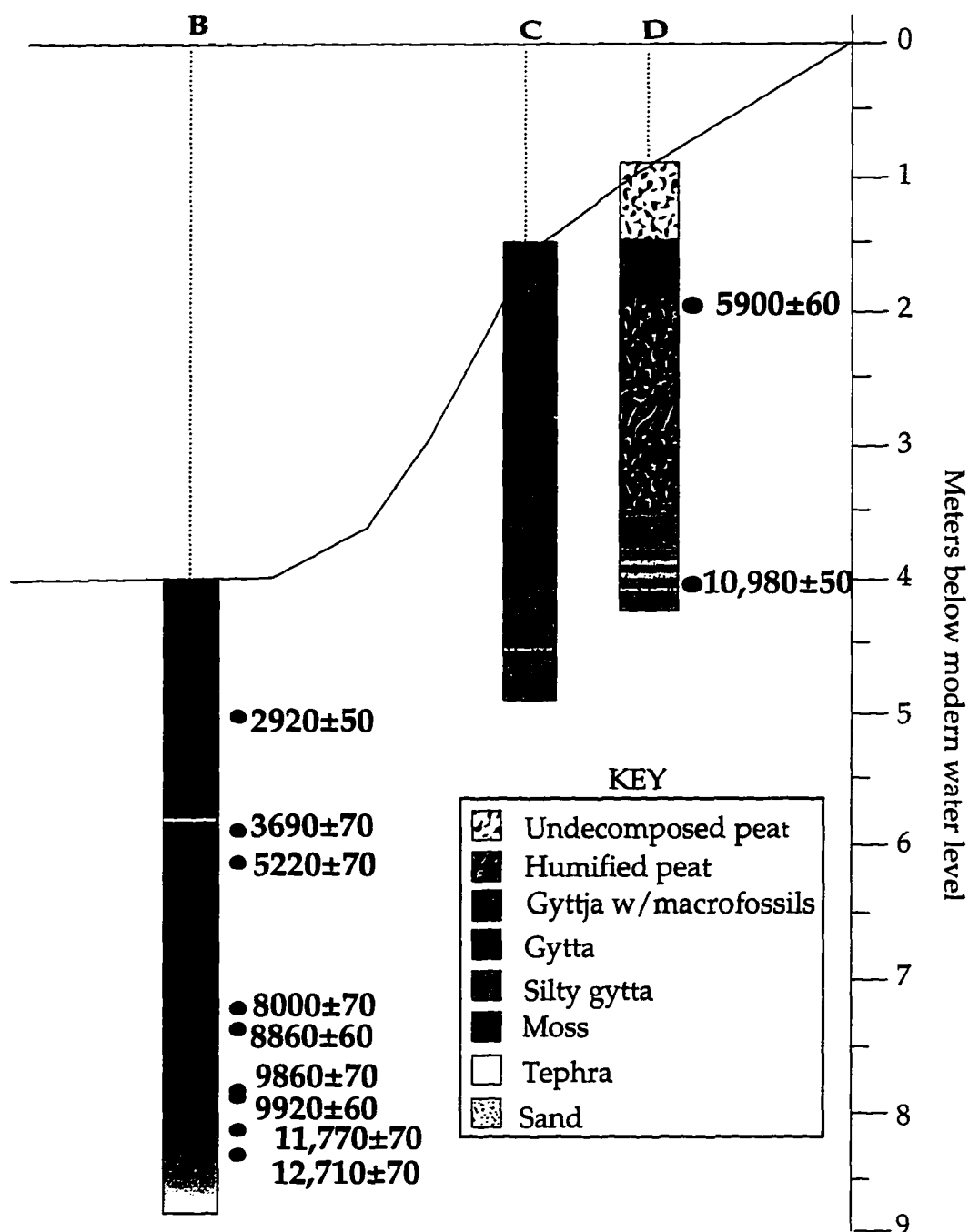


Figure 2.3: Windmill Lake transect stratigraphy

Ash) lies at approximately 580 cm, and three distinct mossy layers lie deeper in the core, approximately at 720 cm, 770 cm, and 780 cm.

Windmill C is dominated by gyttja and abundant macrophyte fragments, with a moss-rich zone between 356 cm and 359 cm. Siltiness gradually increases at about 410 cm; at the core base the sediments are inorganic silt and sand (Figure 2.3).

In Windmill D, the core top is composed of fibrous peat, which gradually changes to mixed gyttja and macrofossils at about 145 cm. A band of gyttja is present between 182 cm and 193 cm; beneath this is a peat with abundant *Carex* and moss fragments. At about 215 cm, the peat gradually becomes more humified, which extends to about 370 cm. A moss layer at about 305 cm interrupts the humified peat. Silty gyttja with lenses of sand dominate the basal 40 to 50 cm of the core (Figure 2.3). The Jarvis Creek Ash was not recognized either visually or through magnetic susceptibility in Windmill C or in Windmill D.

Radiocarbon dates and age models. A total of nine AMS dates on Windmill B, two AMS dates on Windmill D, and a single conventional date on living submerged aquatic mosses were processed from Windmill Lake (Tables 2.1 and 2.2). The analysis on the living aquatic mosses yielded a modern date, indicating that today, anyway, the plants are not absorbing old carbon from the lake water.

Table 2.1: Radiocarbon dates from Windmill Lake, core B.

Depth (cm)	Material dated	mgC	Date	Lab No.	Comments
surface	Aquatic mosses	unavailable	113.6±.09% modern	Beta-75516	$^{13}\text{C}/^{12}\text{C}$ Corrected, aquatic carbon
497-503	<i>Picea</i> pollen	1.01	2920±50	CAMS-29560	terrestrial carbon
589	<i>Nuphar polysepalum</i> seed	0.98	3690±70	CAMS-22019	terrestrial carbon
625	Aquatic mosses	1.05	5220±70	CAMS-22020	aquatic carbon
717.5-718.5	Aquatic mosses	1.11	8000±70	CAMS-22021	aquatic carbon
740	<i>Nuphar polysepalum</i> seed	0.94	8860±60	CAMS-18374	terrestrial carbon
771-772	Aquatic mosses	1.11	9860±70	CAMS-18622	aquatic carbon
778-779	Aquatic mosses	0.90	9920±60	CAMS-22025	aquatic carbon
810-814	Aquatic mosses	0.46	11,770±70	CAMS-22026	aquatic carbon
828-830	<i>Potamogeton</i> seeds	0.46	12,710±70	CAMS-28647	aquatic and/or terrestrial carbon

Table 2.2: Radiocarbon dates from Windmill Lake, core D.

Depth (cm)	Material dated	mgC	Date	Lab No.	Comments
193-194	<i>Nuphar polysepalum</i> seed	0.92	5900±60	CAMS-28645	terrestrial carbon
397-400	<i>Carex</i> seeds	0.83	10,980±50	CAMS-28646	terrestrial carbon

Several of the dates are on *Nuphar polysepalum* seeds. These seeds are good macrofossils for radiocarbon analysis because *Nuphar* uses atmospheric CO_2 and because the seeds originated from the lake itself, and there is no possibility they represent reworked older material that washed in from the catchment. In Windmill B, the *Nuphar* date at 589 cm lies 5 cm below the Jarvis Creek Ash. This date (3690±70) is in very good agreement with other published dates for this tephra layer (Begét et al., 1991).

Aquatic mosses in Windmill B are a mixture of *Scorpidium scorpioides* and *Drepanocladus crassicosatus* (identified by J. Janssens, University of Minnesota), both of which are calciphiles. The possibility exists that the mosses were absorbing old carbon that leached into the lake from the watershed. This is probably not a serious problem because the date on modern aquatic mosses yielded a modern age, indicating they are not absorbing old carbon today.

A 2nd order polynomial regression is the age model for Windmill B. This regression fits the radiocarbon dates well ($r^2 = 0.990$), with an average residual of 238 years (Figure 2.4). This age model also confirms that the increased organic content near the core top resulted in a more rapid sedimentation rate than in the sand- and silt-rich core base.

An age model for Windmill D was not constructed because only two samples were analyzed for radiocarbon, and the stratigraphic changes in the core suggest the sedimentation rate has changed markedly, thus a regression between the two dates is likely to be wildly inaccurate.

The radiocarbon dates were also calibrated to calendar years (CALIB Rev 3.0, Stuiver and Reimer, 1993). Calibration increases the age of the lower part of the core by as much as 2000 yr, while the age of the upper sediments increases by only about 500 yr (Tables 2.3 and 2.4, Figure 2.4). One consequence is a slower estimated sedimentation rate at the base of Windmill B because calibration increases the time interval between samples.

Table 2.3: Windmill B calibrated ages.

Uncalibrated date	Intercepts BP	Mean Intercept	1 S.D. calibrated range (BP)
2920±50	3065	3065	2958-3146°
3690±70	3990, 4040, 4060	4030	3903-4129°
5220±70	5940, 5980	5960	5915-6164*
8000±70	8770, 8790, 8950	8840	8659-8984°
8860±60	9890	9890	9696-9929†
9860±60	11,000	11,000	10,987-11,072†
9920±60	11,010	11,010	10,988-11,202*
11,770±70	13,720	13,720	13,598-13,855*
12,710±70	14,970	14,970	14,771-15,319*

°Calibrated with Pearson and Stuiver (1993) bidecadal bristle cone pine calibration.

*Calibrated with Stuiver and Pearson (1993) bidecadal bristle cone pine calibration .

°Calibrated with Pearson et al., 1993 and Linick et al., 1993, bristle cone pine and German and Irish oak calibration.

† Calibrated with Kromer and Becker (1993) German oak and pine calibration.

*Calibrated with Bard et al. 1993 U-Th calibration on Barbados corals (spline fitted to oak/pine calibration).

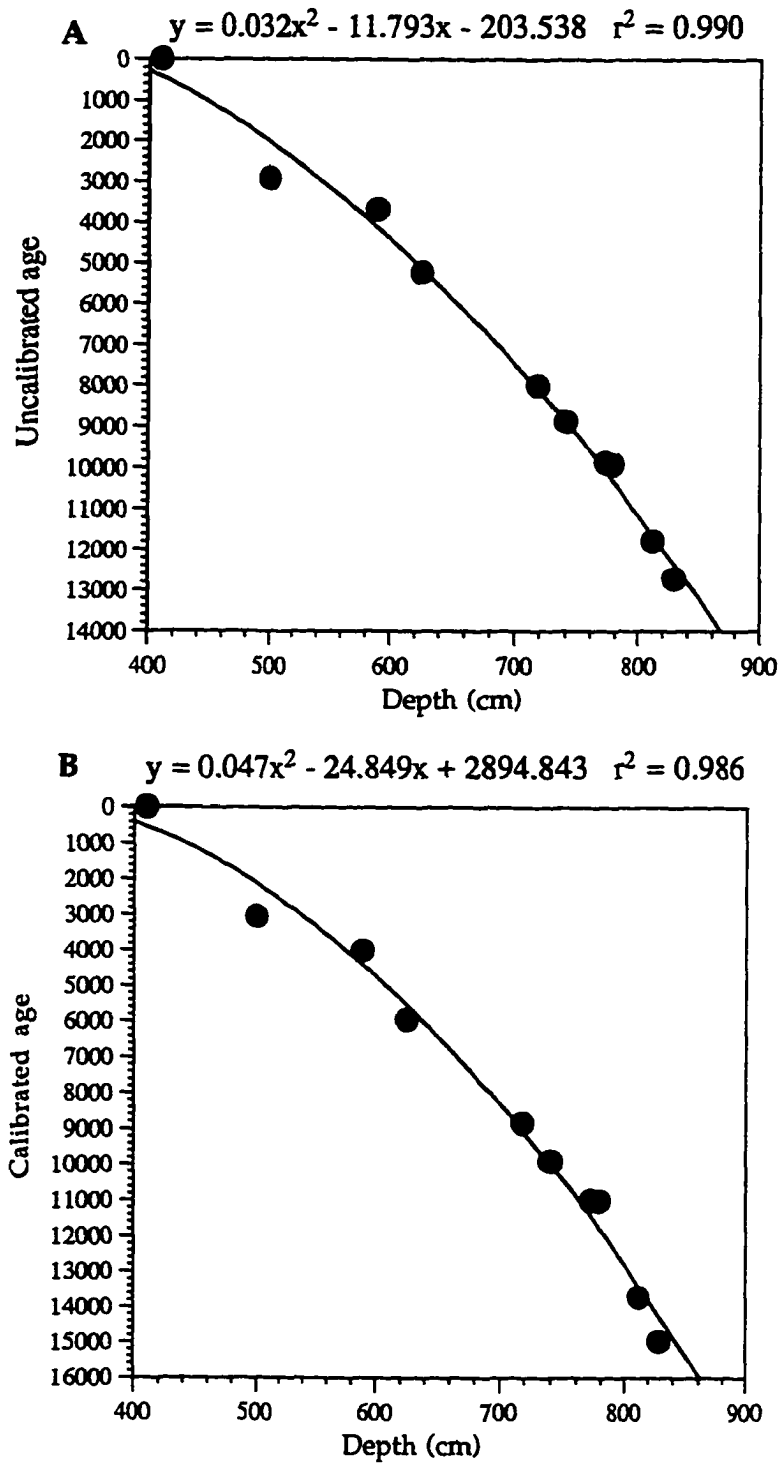


Figure 2.4: Uncalibrated (A) and calibrated (B) age models for Windmill Lake Core A.

Table 2.4: Windmill D calibrated ages.

Uncalibrated date	Intercepts BP	Mean Intercept	1 S.D. calibrated range (BP)
5900±60	6730	6730	6669-6841 ^a
10,980±50	12,898	12,898	12,828-12,965*

Uncalibrated radiocarbon dates and their associated age models are used throughout this study. The calibrated ages are used only when comparing the Windmill lake data with records that are in calendar years, for example, climate model simulations, or when it is thought the uncalibrated time scale may be affecting the results.

The timing of changes in the cores is always made in reference to the age model, not to the specific radiocarbon dates. For example, the AMS date below the Jarvis Creek Ash is about 3700 ¹⁴C yr BP, indicating the tephra was deposited shortly before that time. However, because of curve-fitting associated with the age model, the age of the tephra-fall appears to be about 3800 ¹⁴C yr BP. This kind of discrepancy is a drawback to using a smooth curve to approximate the radiocarbon dates.

Linear interpolation between the dates alleviates this problem. However, linear interpolation introduces unacceptably abrupt shifts in sedimentation rate, and for this reason this method was not used for any of the age models.

A smooth curve also reduces the importance of individual dates, while retaining the overall shape of the age/depth relationship. A smooth curve may also smooth out real changes in sedimentation rate. This problem can be reduced, however, by studying the residuals of the individual dates, and making sure the curve lies within several hundred years of each date. If a single date is a significant outlier, then one must decide whether the date is accurate, and whether there is independent sedimentological evidence to support a large change in sedimentation rate.

Loss on Ignition (LOI) and Magnetic Susceptibility. LOI and magnetic susceptibility reflect the changing organic content of Windmill B (Figure 2.5). LOI at 500°C measures organic carbon, while LOI at 850°C measures inorganic carbon (Bengtsson and Enell, 1986; Dean, 1974).

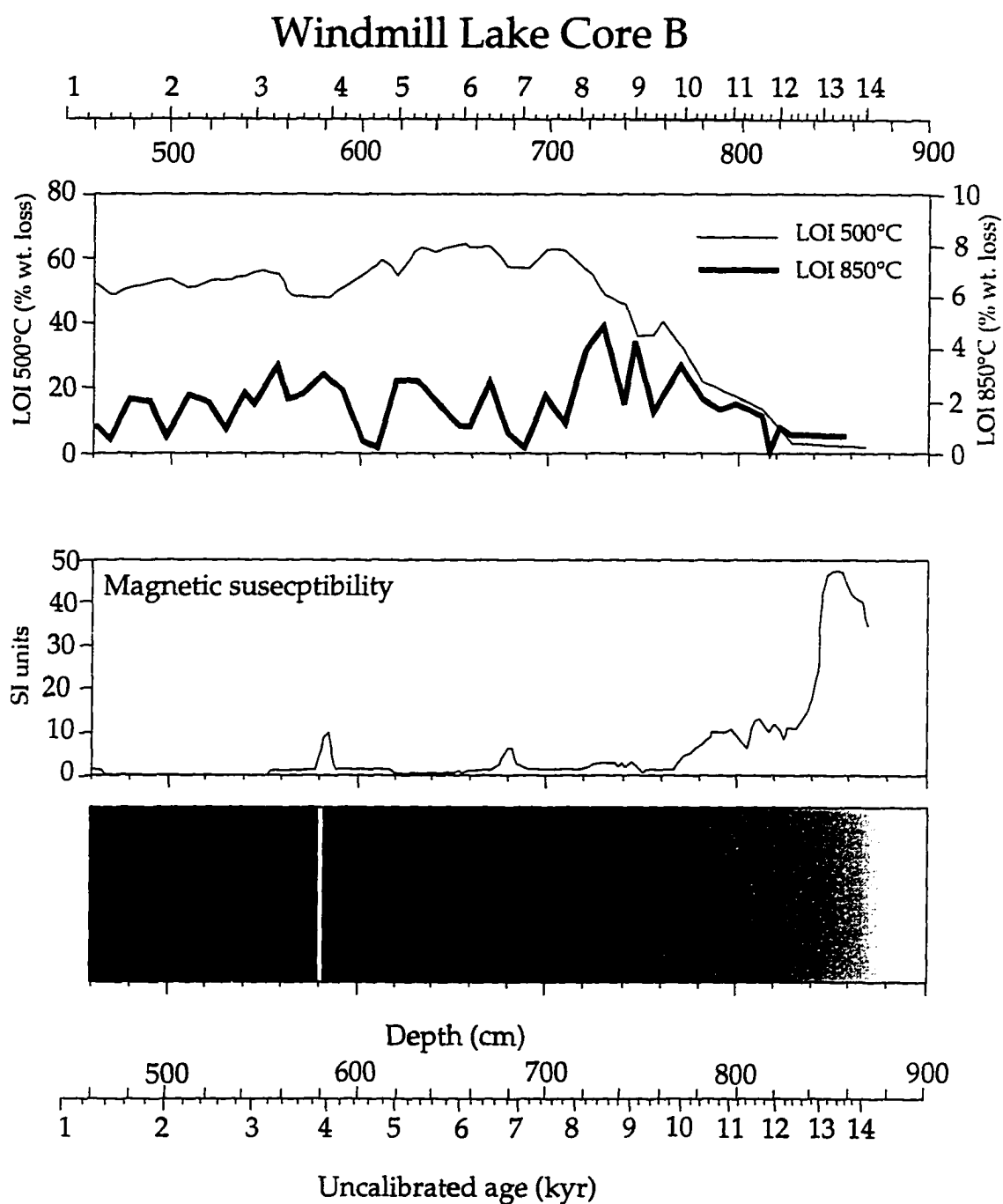


Figure 2.5: Windmill B stratigraphy, LOI, and magnetic susceptibility. See Figure 2.3 for key to stratigraphic patterns.

At the core base, magnetic susceptibility is quite high, but decreases rapidly about 13,000 ^{14}C yr BP. At this time, LOI (500°C) remains low. From about 12,000 to about 10,500 ^{14}C yr BP, magnetic susceptibility is somewhat stable, although LOI (500°C) increases. By about 9800 ^{14}C yr BP, magnetic susceptibility decreases to near 0, although LOI (500°C) continues to increase, reaching modern levels by about 7800 ^{14}C yr BP. Subsequently, the magnetic susceptibility records two peaks, one at about 6800 ^{14}C yr BP, the other about 3800 ^{14}C yr BP. The latter peak is due to the Jarvis Creek Ash. The former peak may also reflect a diffuse tephra layer, although it was not recognized visually or with smear slides.

LOI (850°C) is low throughout the core, never exceeding about 4%. Tests with dilute HCl indicate only the basal sand reacts to the acid. The apparent 4% weight loss between 10,000 and 8000 ^{14}C yr BP may reflect additional water/volatile loss at the higher temperature (Dean, 1974), or the presence of photosynthetic bicarbonate, which may be too fine and/or diffuse to react with the HCl.

Pollen zones. Pollen data are zoned into groups mainly for ease in recognizing contiguous groups of similar samples. In this analysis, visual examination of the pollen diagrams and computerized zoning algorithms were used together to construct the pollen zones.

The computerized zoning algorithms, however, tend to generate a large number of trivial zones. To discover at what point the computerized zoning becomes trivial, one compares the zoning of the actual dataset with the same dataset that has been randomized (see Appendix A for a complete discussion).

Comparison of computer-generated pollen zones for the actual and randomized datasets indicate there are nine (agglomerative [CONISS]) to seven or eight (divisive [SPLITLSQ, SPLITINF]) significant zones in Windmill B (Figure 2.6).

In all the computerized results, zones with less than 4 samples were ignored. This includes the zones marked by boundaries at 12,600, 12,000, 11,700, 3900, and 3600 ^{14}C yr BP. The CONISS and SPLITLSQ results also recognized a boundary at about 1200 ^{14}C yr BP (Figure 2.7). This boundary was ignored because it marks the connection between the surface core and the

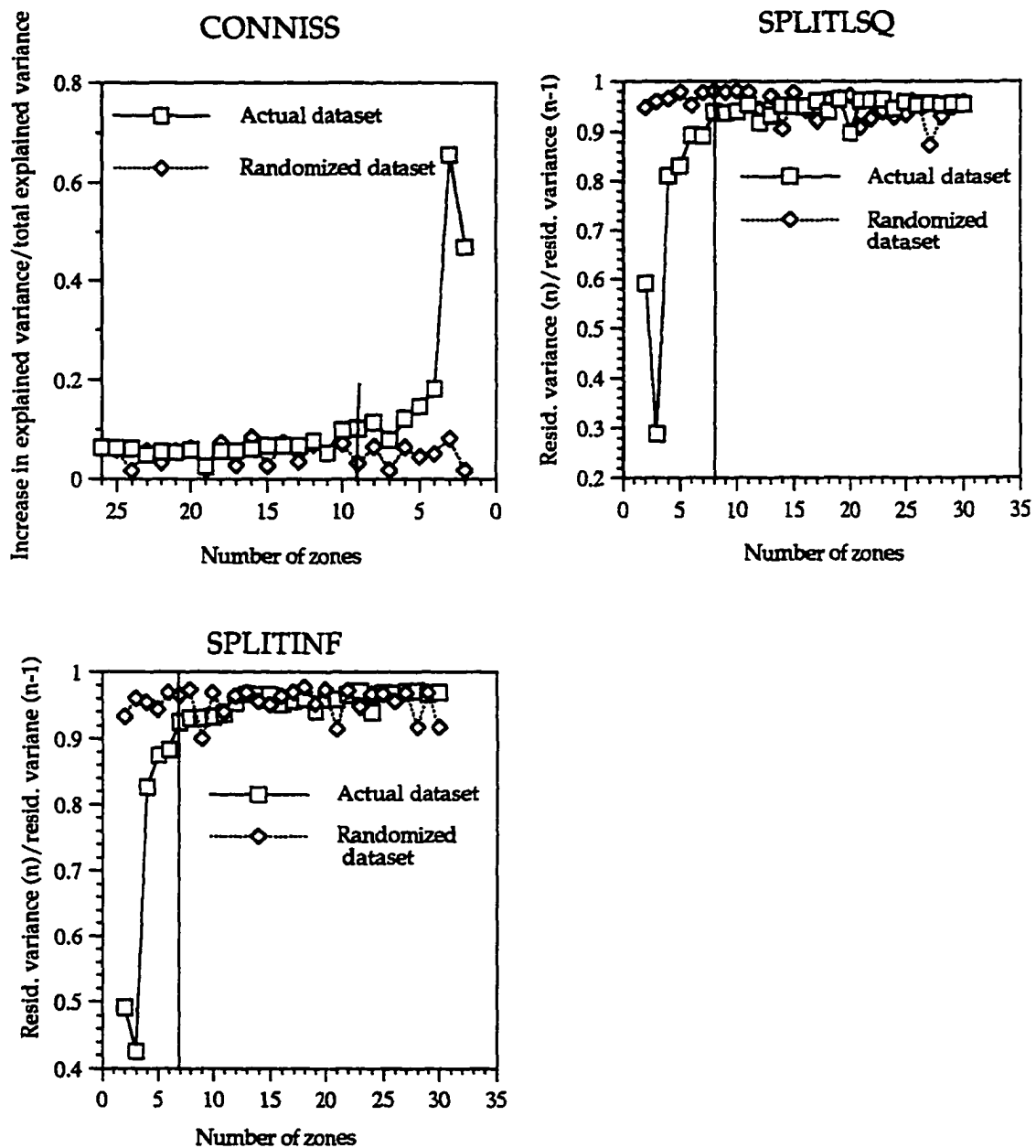


Figure 2.6: Windmill core B, zoning of actual and randomized datasets, lines mark where datasets converge or diverge.

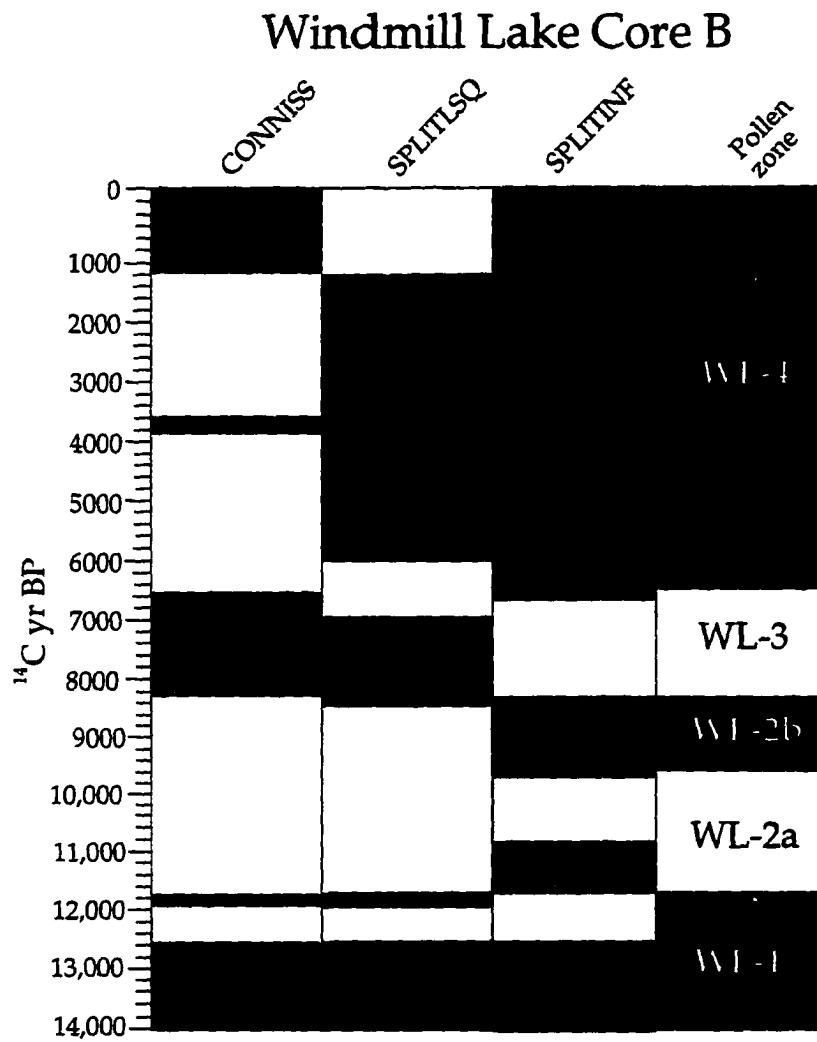


Figure 2.7: Windmill Lake core B pollen zones.

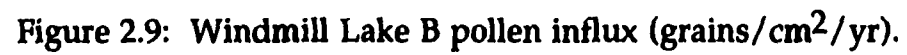
top of the Livingstone core. The abrupt change in the pollen at this time may reflect a previously unrecognized gap in the two core segments.

The zones used for the following discussion are those found consistently with the computer and that indicate a significant change in the vegetation (Figure 2.8). The boundary at 11,700 ^{14}C yr BP marks the *Betula* rise; the boundary at 9600 ^{14}C yr BP marks the *Populus* rise; the boundary at 8400 ^{14}C yr BP marks the *Picea/Alnus* rise, and the boundary at 6500 ^{14}C yr BP marks the stabilization of the *Picea* and *Alnus* frequencies. The 9600 ^{14}C yr BP boundary marks a subzone because it was identified late in the computer zoning analyses, and because it marks a comparatively small change in the vegetation.

Pollen zone WL-1 (>13,000-11,700 ^{14}C yr BP; 807-815 cm) is characterized by low *Betula* and moderate to high *Salix*, *Artemisia*, Cyperaceae, and Poaceae frequencies (Figure 2.8). Influx analysis indicates that total accumulation was low, about 500-800 grains/cm²/yr. The frequency data indicate there is a slight change at the end of the zone with Poaceae and *Artemisia* decreasing, with *Salix* and Cyperaceae increasing, although concentration data (grains/cm³) show all the taxa increasing, indicating the changes are partly an artifact of the percentage calculation (Figure 2.9).

Subzone WL-2a (11,700-9600 ^{14}C yr BP; 815-763 cm) is characterized by very high frequencies of *Betula*, with reduced, but variable frequencies of *Salix*, *Artemisia*, Cyperaceae, and Poaceae (Figure 2.8). Influx analysis indicates that *Betula* dominated the landscape, with significantly lower quantities of *Salix* and herbaceous taxa. Total accumulation is about four times higher than zone WL-1, with all taxa showing increased accumulation (Figure 2.9). Subzone WL-2a also contains a short episode between about 10,000 and 10,800 ^{14}C yr BP where *Artemisia* and Poaceae frequencies and influx increase, while *Betula* and Cyperaceae decrease.

Subzone WL-2b (9600-8400 ^{14}C yr BP; 763-725 cm) is characterized by high *Betula* with moderate *Salix*, Cyperaceae, and Poaceae frequencies, in addition to the presence of *Populus*. Influx analysis indicates *Betula* and *Salix* decrease during the subzone, while the herbaceous taxa are unchanged (Figures 2.8 and 2.9).



Pollen zones WL-3 and WL-4 are similar in the composition of the dominant pollen taxa; they only differ in relative abundance (Figure 2.8). Pollen zone WL-3 (8300-6500 ^{14}C yr BP; 725-675 cm) is a transition zone between the first appearance of *Picea* and *Alnus* and when these taxa achieve modern frequencies. Pollen zone WL-4 (6500 ^{14}C yr BP-present; 675-410 cm) contains all the modern constituents of the boreal forest, although their frequencies (i.e., *Picea* and *Alnus*) are especially variable compared to other taxa in this zone. A marked break in the pollen data occurs at 1200 ^{14}C yr BP, this is where the Livingstone core joins the surface core. Some of the change in the pollen frequencies may be real, although a gap between the two cores probably explains most of the change.

Rate of change. Rate of change of the pollen percentage data was calculated to see where adjacent samples are most different from each other. Chord distance is the dissimilarity coefficient (DC) used in the analysis because it has a good signal to noise ratio (Jacobsen et al., 1987; Lotter et al., 1992). The DCs were calculated on the taxa listed below (Table 2.5).

Table 2.5: Taxa used in rate of change analyses.

<i>Picea</i> undiff.
<i>P. glauca</i>
<i>P. mariana</i>
<i>Juniperus</i>
<i>Betula</i>
<i>Alnus</i>
<i>Salix</i>
<i>Populus</i>
<i>Ericales</i>
<i>Artemisia</i>
<i>Cyperaceae</i>
<i>Poaceae</i>
<i>Anemone</i> -type
<i>Astragalus</i> -type
<i>Rosaceae</i>
<i>Caryophyllaceae</i>
<i>Brassicaceae</i>

The dissimilarity coefficients indicate episodes where the taxon proportions change in adjacent samples. The highest DCs are found about

11,700, 10,400, 8000, 3000-4000, and about 1200 ^{14}C yr BP. The 11,700- ^{14}C yr peak is due to the transition from herb-dominated pollen to shrub-dominated pollen. The 10,400-yr peak reflects the transient increase in herbs (especially Poaceae) and xeric taxa (*Juniperus* and *Artemisia*) at the expense of mesic taxa (*Betula*). A group of peaks at about 8000 ^{14}C yr BP is due to the appearance of *Picea* and *Alnus* in the pollen core. The later peaks in the Holocene are caused by variations in the relative abundance of *Picea* and *Alnus*.

To calculate the rate of change (per 100 ^{14}C yr), the dataset is interpolated to a uniform sample interval to remove spuriously high rates of change where sample intervals are small. Rate of change on the interpolated dataset (100 yr intervals) emphasizes the dominant peaks and down-weights some of the noise. In this analysis, the 11,700- ^{14}C yr peak continues to dominate, but the later peaks are about equal in size, with a rate of change about 1/3 of the 11,700- ^{14}C yr peak (Figure 2.10).

Rate of change was also calculated using the calibrated age model. This did not significantly alter the analysis. It only increased the age and slightly reduced the dissimilarity of the older peaks, while increasing the dissimilarity of the younger peaks (Figure 2.11).

Ordination. The Windmill B pollen data were ordinated to discover any underlying structure in the dataset. The analysis uses the same taxa as the rate of change analysis, but with a square root transformation. The first axis gradient length is 2.137 standard deviation units, indicating that Principle Components Analysis (PCA) is an appropriate method for this dataset. In the PCA, axis 1 explains 54.4% of the variability, and axis 2 explains an additional 31.7% of the variability. The PCA clearly separates the different pollen zones (Figure 2.12). Samples in pollen zone WL-1 are characterized by low loadings on the first axis, but high loadings on the second axis. The remaining pollen zones are characterized by low to high loadings on the first axis, but consistently low loadings on the second axis.

The first axis is characterized by high loadings of *Picea*, *Alnus*, and Ericales, and low loadings of *Salix* and the herbaceous taxa. The second axis has high loadings in the herbaceous taxa, and low loadings with *Betula* and *Populus* (Table 2.6).

Windmill Lake Core B

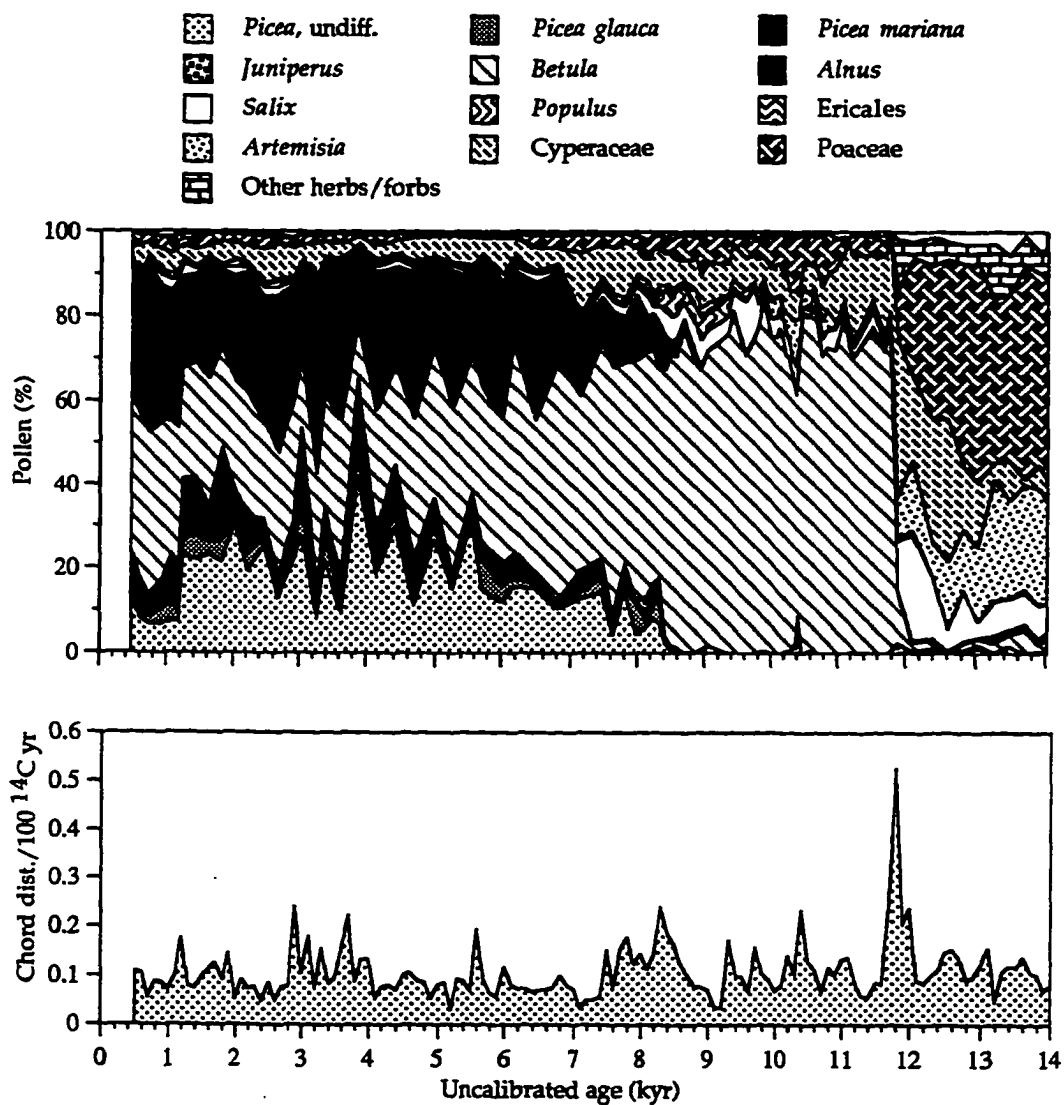


Figure 2.10: Windmill B rate of change; uncalibrated age scale.

Windmill Lake Core B

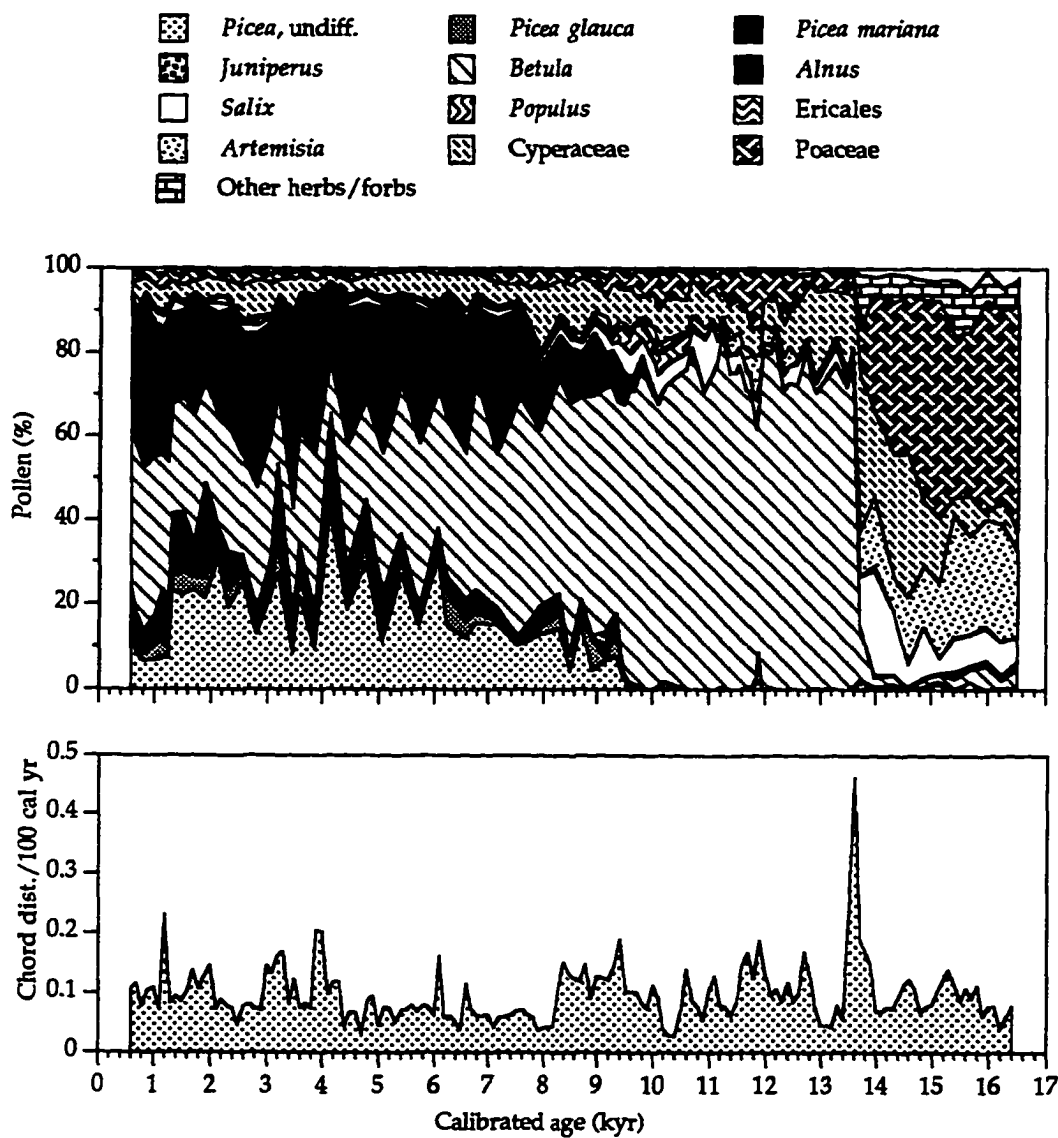
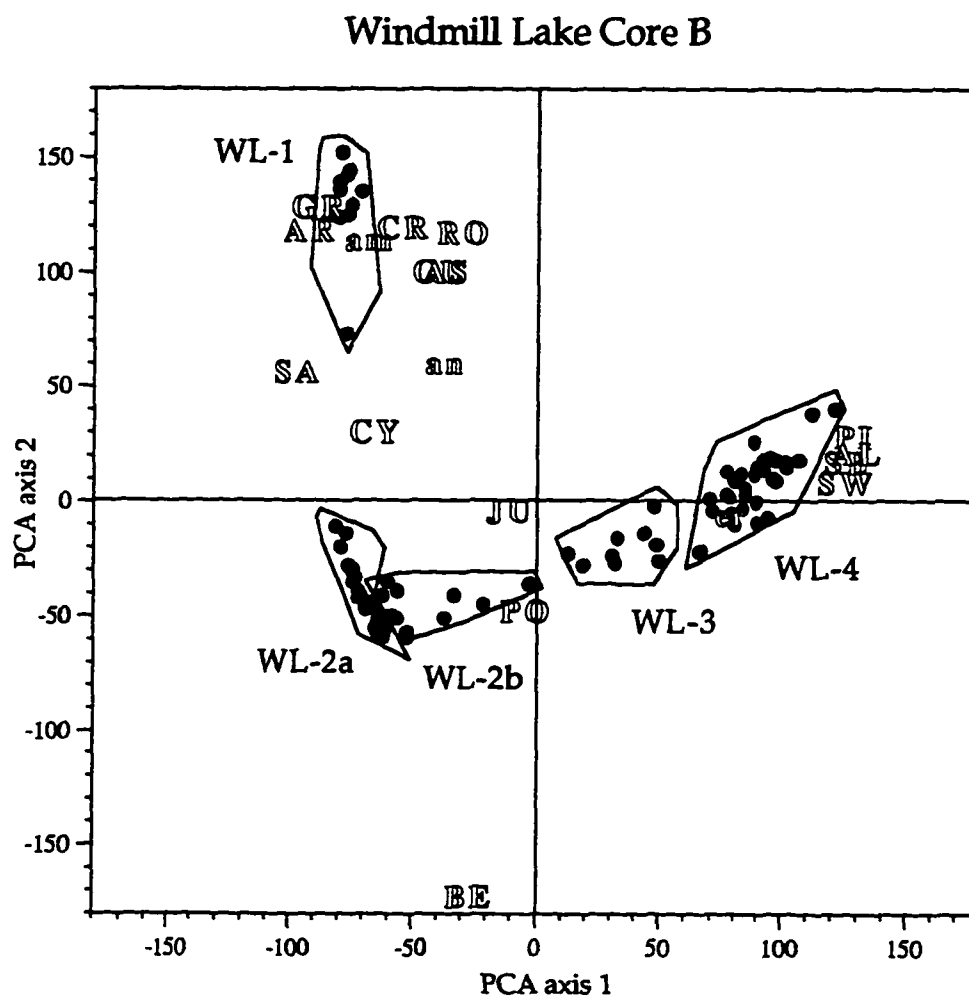


Figure 2.11: Windmill B rate of change, calibrated age scale.



Taxon abbreviations

PI= <i>Picea</i> undiff.	am= <i>Solidago</i> -type
SB= <i>P. mariana</i>	CR=Caryophyllaceae
SW= <i>P. glauca</i>	CU=Brassicaceae
er= <i>Ericales</i> undiff	AS= <i>Astragalus</i> -type
PO= <i>Populus</i>	RO=Rosaceae
JU= <i>Juniperus</i>	an= <i>Anemone</i> -type
CY=Cyperaceae	BE= <i>Betula</i>
SA= <i>Salix</i>	AL= <i>Alnus</i>
GR=Poaceae	
AR= <i>Artemisia</i>	

Figure 2.12: Windmill core B PCA biplot.

Table 2.6: Species loadings on the first two axes.

Species	Axis 1	Axis 2
<i>Picea undiff.</i>	129	27
<i>P. glauca</i>	126	7
<i>P. mariana</i>	127	16
<i>Juniperus</i>	-10	-5
<i>Betula</i>	-26	-174
<i>Alnus</i>	130	20
<i>Salix</i>	-96	55
<i>Populus</i>	-4	-49
Ericales	72	-7
Herbaceous taxa (mean)	-57	97

When the ordination scores are plotted against sample age, useful patterns emerge (Figure 2.13). Most of the variability in the dataset is modeled on axis 1, while axis 2 models the variability only in pollen zones WL-1 and WL-2. **Carbon and Nitrogen concentration and fluxes.** Carbon and nitrogen concentrations, as well as fluxes may provide information on lake productivity and terrestrial input. Carbon and nitrogen concentrations in Windmill B gradually increase up-core, with both elements reaching their maximum amounts about 8000 ^{14}C yr BP. Subsequently, both C and N decrease slightly (Figure 2.14).

Carbon and nitrogen fluxes (in $\text{g}/\text{cm}^2/\text{yr}$) can be used as a proxy for lake productivity, because they estimate the rate C and N of accumulation in the lake basin, if the organic matter in the lacustrine sediments was produced in the lake. However, the fluxes are also sensitive to changes in bulk density, as well as sedimentation rate which may be independent of productivity. C and N fluxes vary in Windmill B, with peaks at 11,700 ^{14}C yr BP, and between 10,600 and 9800 ^{14}C yr BP. Near modern levels are achieved by about 8000 ^{14}C yr BP (Figure 2.14). The increased fluxes at 11,700 and 10,600 to 9800 ^{14}C yr BP are due to increases in bulk density, as well as slight increases in C and N concentrations.

The carbon/nitrogen ratio provides a rough estimate of aquatic and terrestrial inputs into the lacustrine sediments (LaZerte, 1983). The carbon/nitrogen ratio is somewhat variable throughout the core, ranging between about 12 and 17 (Figure 2.14). At the core base, C/N ratios fluctuate

Windmill Lake Core B

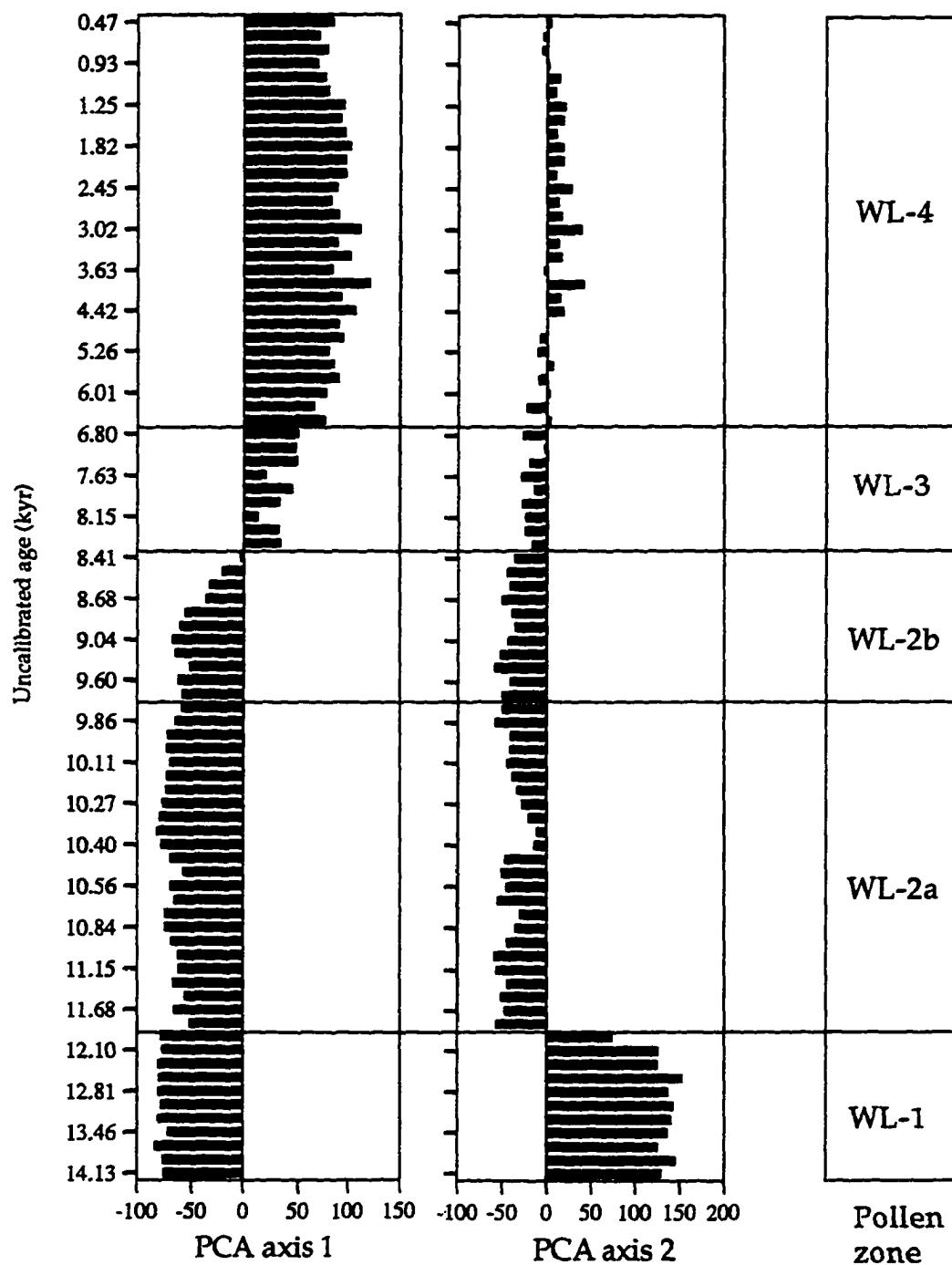


Figure 2.13: Windmill Lake PCA sample scores plotted against age.

Windmill Lake Core B

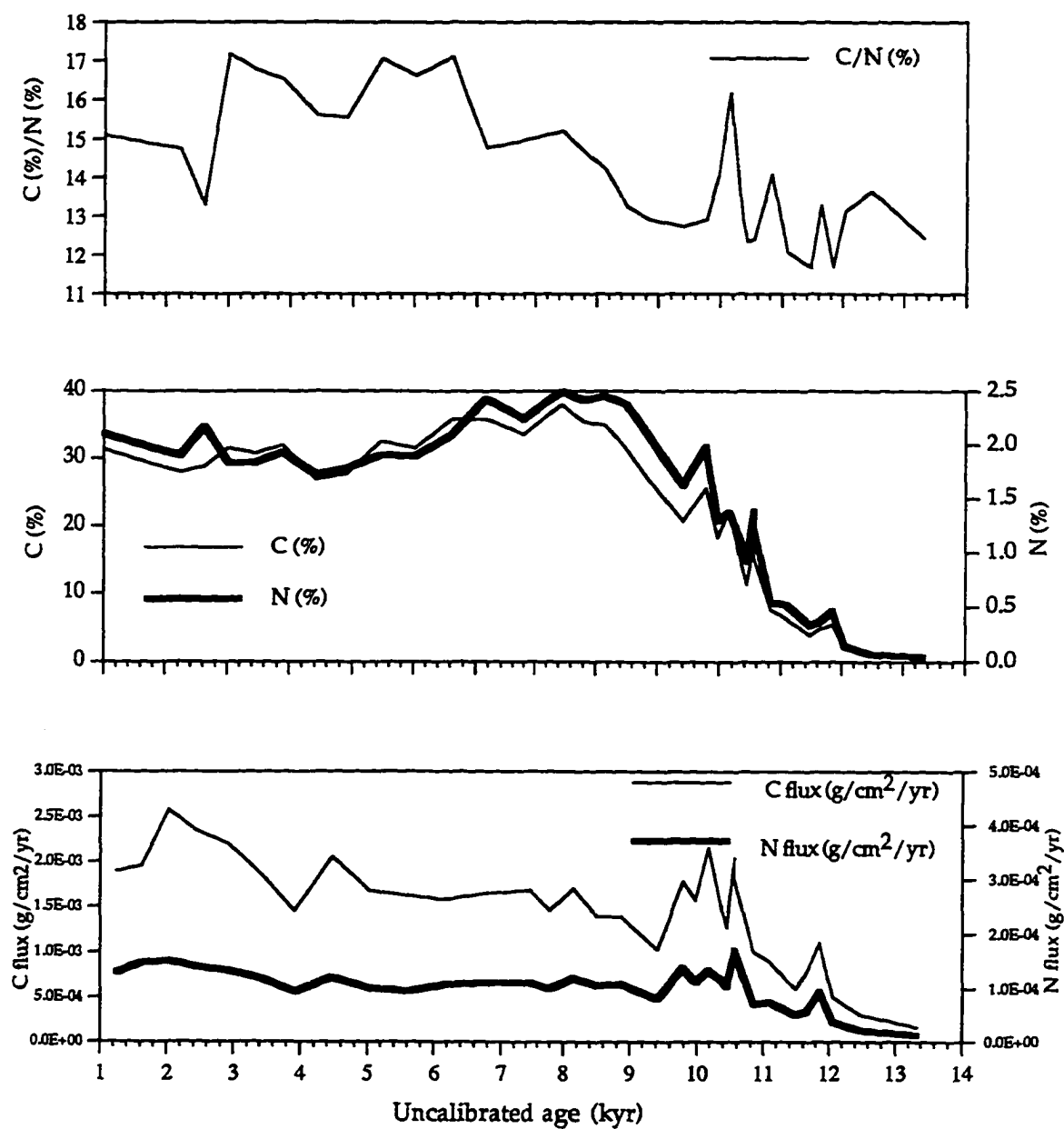


Figure 2.14: Windmill core B, carbon (%), nitrogen (%), C/N ratio, and carbon and nitrogen fluxes (g/cm²/yr).

around 12, with a sharp spike of 16 at about 10,200 ^{14}C yr BP. The C/N ratio subsequently decreases to about 12, but then gradually increases at about 8400 ^{14}C yr BP, reaching a maximum value of about 17 at about 5600 ^{14}C yr BP.

Carbon and Nitrogen Isotopes. The $\delta^{13}\text{C}$ values of sediments in the Windmill B core range between about -31‰ and -24‰ (Figure 2.15). The base of Windmill core B is relatively enriched in $\delta^{13}\text{C}$, initially decreasing rapidly about 11,700 ^{14}C yr BP, with a subsequent step-wise decrease at about 10,700 ^{14}C yr BP. This is followed by a brief increase in $\delta^{13}\text{C}$ at about 10,500 ^{14}C yr BP. Subsequently, $\delta^{13}\text{C}$ values decrease again, and remain fairly constant until about 6200 ^{14}C yr BP. At this time $\delta^{13}\text{C}$ values increase markedly to a stable level until about 2000 ^{14}C yr BP, at which time $\delta^{13}\text{C}$ decreases again to pre-6000 ^{14}C yr levels.

At Windmill lake, the $\delta^{15}\text{N}$ of the sediments ranges between about 0.5‰ and 1.7‰. At the core base (Figure 2.15), $\delta^{15}\text{N}$ is about +1.5‰, a brief increase to >+2‰ occurs between about 10,600 ^{14}C yr and 9800 ^{14}C yr BP. Subsequently the $\delta^{15}\text{N}$ falls to 1.5‰, and then to 0.5‰. A short increase to about +1‰ occurs between 7800 and 6800 ^{14}C yr BP. Beginning about 6000 ^{14}C yr BP and continuing to the present, the $\delta^{15}\text{N}$ gradually increases from <+1‰ to about +1.6‰.

Discussion

Vegetation reconstruction. The pollen data from pollen zone WL-1 (herb zone) (Figure 2.8) suggest the regional vegetation was dominated by herbaceous taxa that sparsely covered the landscape. The pollen at this time is dominated by the "big three", i.e., Poaceae, Cyperaceae, and *Artemisia*. Combined, these taxa make up about 90% of the total pollen. The remaining 10% consists of *Salix* and forbs, primarily Rosaceae, Caryophyllaceae, and Asteraceae (mainly Tubuliflorae). The mesophytic taxa (*Salix*, most Cyperaceae, and Rosaceae) would have grown in damper sites, such as near the lake margin, or on the hillsides as snow-bed communities. The xerophytic taxa (*Artemisia* and some Poaceae) would have grown in well-drained, or wind-swept sites that had a thin snow cover.

Windmill Lake Core B

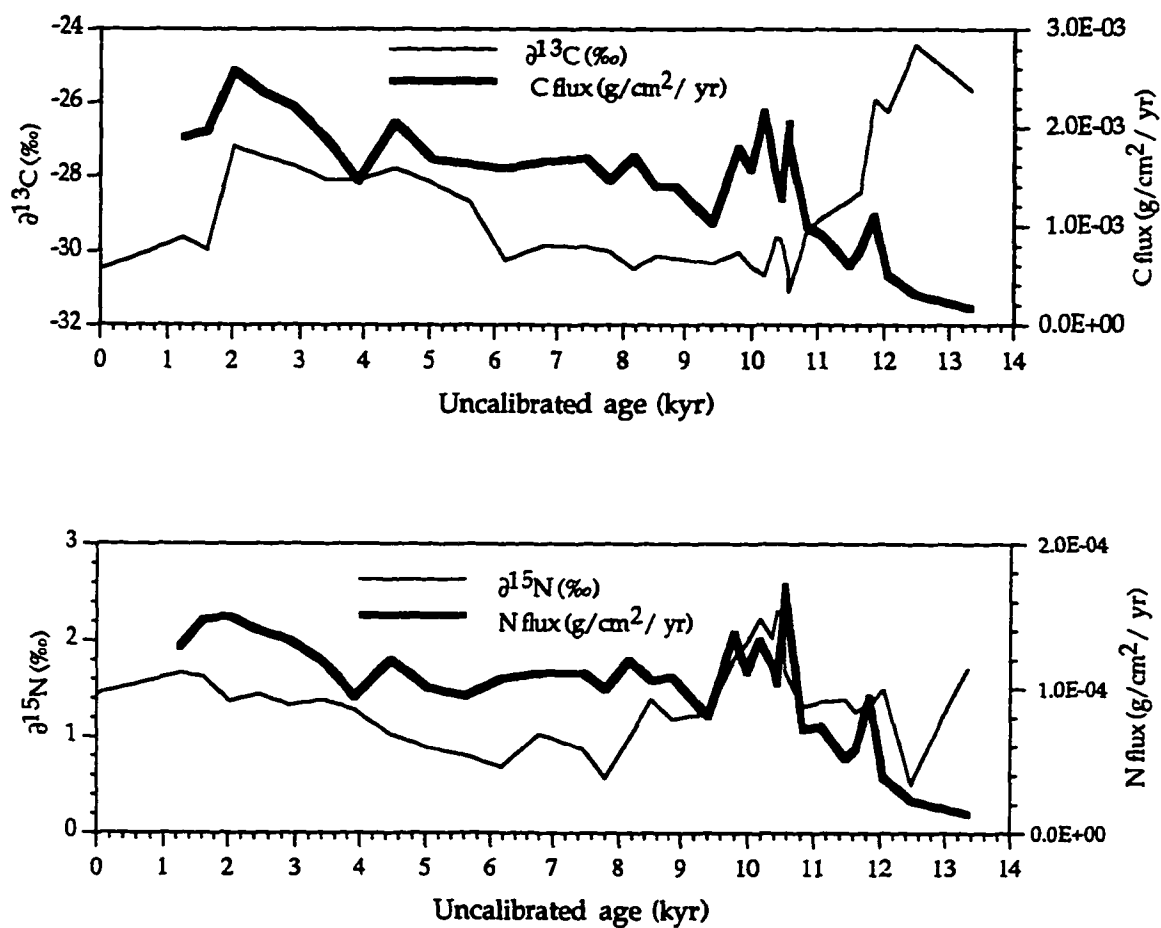


Figure 2.15: Windmill core B, carbon and nitrogen isotopes, as well as fluxes.

These herb zone taxa form a unique assemblage that, while different from most modern plant communities, is most similar to the high Arctic tundra vegetation (Anderson et al., 1989). As in most lake records where the herb zone is preserved (c.f. Lateral Pond [Cwynar, 1982]; Hanging Lake [Ritchie, 1982]; Tiinkdhul [Anderson et al., 1988]; Screaming Yellowlegs Pond [Edwards et al., 1985]; Niliq [Anderson, 1988]; Kaiyak and Squirrel lakes [Anderson, 1985]; Joe Lake [Anderson, 1987; Anderson et al., 1994]), the herb zone total terrestrial pollen influx at Windmill Lake is very low, less than 1/10th of later zones, suggesting a sparse vegetation cover.

At Windmill Lake, the appearance and dominance of *Betula* pollen occurred over a period of less than 100 ^{14}C years, at about 11,700 ^{14}C yr BP. Other pollen records from eastern central Alaska (Birch, McFarlane, Tiinkdhul) and the Yukon Territory (Hanging Lake, Lateral Pond) (Figure 1.2), also have a very abrupt birch rise, indicating the transition occurred rapidly over a large area (Ager, 1975; Anderson et al., 1988; Cwynar, 1982, Ritchie, 1982). This suggests that *Betula* was probably already present throughout the region in scattered localities, possibly only reproducing vegetatively.

Influx calculations (Figure 2.9) also suggest that while *Betula* became an important taxon at Windmill Lake, the herbaceous taxa from the preceding period continued, often at greater quantities than during the herb zone. In modern pollen rain studies of the northern boreal forest, *Betula* pollen over-represents the vegetation (Ritchie, 1977; Ritchie, 1982; Short et al., 1986). Despite the dramatic appearance of the *Betula* pollen, landscape coverage may have been relatively low, and the herbaceous taxa continued and probably expanded. Data on modern pollen/vegetation relationships (Ritchie, 1982) in the Yukon Territory indicate that the area of *Betula* vegetation may have been small, and that the significant vegetation change was a general increase in plant coverage on the landscape.

While macrofossil analysis has not been conducted, most of the *Betula* pollen is probably from dwarf (*Betula nana*) and shrubby (*Betula glandulosa*) varieties, because tree-birch is generally not present until later in the Holocene (Hu et al., 1993). Birches are mesic plants, with *B. glandulosa* and *B.*

nana occupying lowlands and poorly drained areas in the boreal forest, as well as mesic alpine slopes. Both species are also present in poorly drained shrub- and tussock tundra communities beyond latitudinal treeline (Hultén, 1968; Viereck et al., 1992). At the time of the birch rise, they probably occupied similar topographic settings as today, leaving the drier and disturbed areas to the more xerophytic weedy species such as *Artemisia*, Poaceae, Cyperaceae, Asteraceae, Brassicaceae, and Chenopodiaceae-Amaranthaceae.

The timing of this vegetation change is also important. The data at Windmill Lake suggest the *Betula* rise occurred about 11,800 ^{14}C yr BP. Conventional bulk radiocarbon dates from lakes in the Tanana valley (Ager, 1983), in addition to the eastern and central interior, date the birch rise to about 13,000-14,000 ^{14}C yr BP (Anderson et al., 1994; Anderson et al., 1988; Cwynar, 1982; Edwards et al., 1985). The 11,800 ^{14}C yr date from Windmill Lake agrees well with a new AMS date from Birch Lake (discussed in Chapter 4), suggesting the Windmill date is not anomalous. The previously reported dates are, without exception, conventional dates on bulk organic matter. As the use of AMS technology becomes more widespread, it is becoming apparent that conventional dates are inherently unreliable, because they amalgamate carbon from unknown sources which may or may not be in equilibrium with the atmosphere (Abbott and Stafford, 1996).

A younger age for the birch rise is important primarily because it coincides with the initial peopling of the New World. The earliest well-dated archaeological sites in Alaska date to about 11,700 ^{14}C yr BP (Dry Creek, Walker Road, Broken Mammoth, Mesa [Hoffecker et al., 1993; Powers and Hoffecker, 1989; Goebel et al., 1991; Yesner et al., 1992; Kunz and Reanier, 1994]) (Figures 1.3 and 2.1). A direct connection between climatic amelioration and the arrival of humans is difficult to prove. However, a warmer and wetter climate probably made it easier for people to overwinter in high latitudes. This topic is discussed further in Chapter 6.

Between about 9500 and 8500 ^{14}C yr BP, *Populus* pollen is present in Windmill B. *Populus* pollen (*P. balsamifera*) does not travel far from its source, nor does it preserve well (Edwards and Dunwiddie, 1985; Sangster and

Dale, 1961), implying that its presence in a pollen core indicates the trees grew locally, possibly at the lake margin.

It is unclear whether the *Populus* pollen represents *P. balsamifera* or *P. tremuloides*. *Populus* pollen has only occasionally been identified to the species level (Brubaker et al., 1983), suggesting that in the central Brooks Range, anyway, *P. balsamifera* was the dominant taxon in the early Holocene. *P. balsamifera* leaves dated to about 13,700 ^{14}C yr BP have also been found far north of its current range (Hopkins et al., 1981), suggesting the species may have been widespread during the early Holocene. *P. balsamifera* is a shade intolerant pioneering species, commonly found on river floodplains and other well-drained sites (Viereck et al., 1992). It is also highly sensitive to summer temperatures, where growth rates (for both treeline and non-treeline stands) are positively correlated with early summer warmth (Lev, 1987; Edwards and Dunwiddie, 1985). This suggests that if the *Populus* pollen at Windmill Lake represents *P. balsamifera*, then the expansion of this taxon may indicate increasing summer warmth, or perhaps extensive floodplain development in the Nenana valley.

The timing of the *Populus* subzone at Windmill Lake agrees fairly well with other nearby lakes such as Eightmile and Birch lakes (Ager, 1983) (Figure 2.1), but precise chronological comparisons are often impossible because of distant limiting ages and contamination of the radiocarbon samples by old carbon (Bartlein et al., 1995).

Wherever the *Populus* subzone has been identified, it usually dates between about 11,000 and 8000 ^{14}C yr BP (Bartlein et al., 1995). At all sites, the *Populus* subzone ends before the local arrival of spruce. The *Populus* subzone is widely distributed throughout most of Alaska, being absent or much reduced in high elevation and coastal sites. The sites with the highest percentage of *Populus* ($\geq 25\%$ of pollen sum) are at low elevations, often in riverine settings, such as Ruppert, Wien, M Lake, and Farewell lakes (Brubaker et al., 1983; Hu et al., 1993; Ritchie, 1977; Hu et al., 1996) (Figure 1.2).

A number of sites show a *Salix* peak during the *Populus* subzone, suggesting that the *Populus* and *Salix* may have formed a scrub assemblage

either on south-facing slopes (Ager, 1983), or in riverine settings (Anderson et al., 1988).

At Windmill Lake, the absence of a strong *Salix* peak, and the moderately low percentages of *Populus* (ca 5%), suggest that these taxa did not form a widespread community at this locality. *Populus* is not present in the watershed today, although some stunted trees (*P. tremuloides*) grow on the southwest-facing slopes that separate the Windmill watershed from the Nenana River.

At about 8400 ^{14}C yr BP, the arrival of *Picea* and *Alnus* at Windmill Lake marks a significant change in the local vegetation and ecosystem. Modern pollen/vegetation relationships suggest that 10% *Picea* pollen indicates the tree was present locally (Anderson and Brubaker, 1986), although this figure may be closer to 5% (Hu et al., 1993). At Windmill Lake the 10% *Picea* threshold is crossed at about 8300 ^{14}C yr BP, suggesting the trees were present in the Windmill Lake watershed at this time. *Alnus* pollen, while becoming more common with the arrival of *Picea*, does not cross the 20% *Alnus* threshold (indicating local presence [Anderson and Brubaker, 1986]), until significantly later, about 6800 ^{14}C yr BP. This 1500 ^{14}C yr period between the local arrivals of *Picea* and *Alnus* marks a transition period from a shrubby, *Betula*-dominated landscape with scattered *Populus* groves and spruce trees to a closed boreal forest somewhat similar to the modern vegetation.

The ultimate source area for the *Picea* is probably southern Canada. After deglaciation, *Picea* rapidly migrated from source areas in southwestern Canada, northwards, to the southern Yukon, reaching eastern Alaska and the Mackenzie delta by 9000 ^{14}C yr BP (Hopkins et al., 1981; Ritchie, 1984; Ritchie and MacDonald, 1985). In northern Canada, *Picea* extended far beyond the modern treeline at this time, suggesting significant climatic warmth, probably warmer than today (Ritchie et al., 1983).

Inside Alaska, *Picea* continued migrating after the initial arrival, although the westward and northward movement had virtually ceased by about 6000 ^{14}C yr BP (Anderson and Brubaker, 1994). Traditionally, the earliest arrival of *Picea* is dated to roughly between 9400 ^{14}C yr BP (spruce cone in an eolian deposit in the Tanana valley [Weber et al., 1981]) and 9000 ^{14}C yr BP (pollen

records from the Tanana valley [Ager, 1975, 1983]). However, a new AMS date from one of the Tanana valley pollen records (Birch Lake), suggest *Picea* did not actually arrive until about 8400 to 8000 ^{14}C yr BP (see Chapter 4).

This change has two implications. First, the new dates indicate that *Picea* migration took place simultaneously in the Tanana and Porcupine/Upper Yukon valleys (Anderson et al., 1988; Edwards and Brubaker, 1986). Second, *Picea* virtually exploded across central Alaska from east to west, showing only a slight delay over the 400 km between Birch and Farewell lakes, but then nearly halted, arriving at Bristol Bay only 3000 ^{14}C yr BP (Table 2.7).

Table 2.7: Arrival times of *Picea* at various lakes in central Alaska.

Site (east to west)	^{14}C age at 10% <i>Picea</i>	Reference
Birch Lake (Tanana valley)	8400	Edwards et al. in prep.
Ten Mile (Delta valley)	9000†	Anderson et al, 1994
Eightmile (Nenana valley)	7400†	Ager, 1983
Windmill (Nenana valley)	8300	this pub.
Dune (Tanana valley)	8700	this pub.
Wien (Kantishna valley)	8500	Hu et al., 1993
Farewell (Kuskokwim valley)	8000	Hu et al., 1996
Grandfather (Bristol Bay)	3000	Hu et al., 1995

†Conventional dates on bulk sediments. The Ten Mile date may be too old for this reason. All other dates are AMS.

At Windmill Lake, *P. mariana* was probably present in the region as early as 8300 ^{14}C yr BP, although influx analysis suggests it did not become an important part of the vegetation until after 5200 ^{14}C yr BP (Figure 2.9). Prior to that time, *P. glauca* was the dominant conifer. *P. glauca* probably grew (as did *B. papyrifera* and presumably *P. tremuloides*) on warm, moderately well-drained sites, such as hillsides and stabilized flood plains. The increase in *P. mariana* influx after 5200 ^{14}C yr BP may reflect some paludification of the lowlands, which would have favored *P. mariana* over *P. glauca*, although the latter would have continued to grow on the hillsides.

At about 6800 ^{14}C yr BP, *Alnus* appeared in the Windmill Lake watershed. Low quantities of *Alnus* pollen are present at 9000 ^{14}C yr BP, suggesting scattered clumps grew in the region, but it was not present locally. In Alaska,

Alnus is present earlier and more abundantly in the western and southwestern part of the state (Anderson and Brubaker, 1994; Hu et al., 1995), suggesting these regions (or the Bering Land Bridge) may have been refugia during the full glacial. With the arrival of *Alnus*, the final major species of the modern boreal forest is present.

Development of the modern boreal forest. The *Picea* pollen data show that both percentage and influx values increase (albeit with fluctuations) from its initial arrival, until about 3800 ^{14}C yr BP, when *Picea* percentage and influx were significantly higher than modern values (Figures 2.8 and 2.9). After this time, *Picea* declined to modern values. This pattern is seen at other central Alaskan pollen sites, such as Birch Lake, Dune Lake, Wien Lake, and Wonder Lake (Figure 1.2) although the timing of *Picea* decline is somewhat variable (Hu et al., 1993; Anderson et al., 1994). This suggests that during the mid-Holocene, *Picea* grew more densely or was more widely distributed across the landscape, and that after about 4000 ^{14}C yr BP, climatic conditions changed and *Picea* became less common.

It is unclear whether the *Picea* decline at Windmill Lake represents a reduction of *P. glauca* or *P. mariana*, because most of the change is in the undifferentiated category. The *Picea* decline may be due to climatic deterioration. By 4000 ^{14}C yr BP, various proxy climatic records indicate the onset of climatic cooling. These include the retraction of treeline in northwest Canada (Ritchie, et al., 1983; Moser and MacDonald, 1990) and the expansion of alpine glaciers in Alaska (Hamilton, 1986b; Calkin, 1988). Even a slight reduction of summer warmth could have triggered significant landscape changes. In the uplands, particularly near treeline, *P. glauca* may have been restricted to the warmest sites. In the lowlands, with *P. mariana* already abundant, cooler summers could have initiated paludification and permafrost growth, suggesting that the modern boreal forest, with its extensive paludification and widespread permafrost probably did not form around Windmill Lake until after 4000 ^{14}C yr BP.

Variability in the abundance of boreal forest taxa. Between about 5800 and 3000 ^{14}C yr BP, the relative abundances of the *Picea*, *Betula* and *Alnus* at Windmill Lake fluctuate markedly (Figure 2.8). The fluctuations are most

dramatic in the percentage data, although they are still present in the influx data, which suggests they reflect real changes in the vegetation (Figure 2.9). In both calculations *Betula* and *Alnus* change together, but usually opposite to the *Picea* fluctuations. Unfortunately, most of the fluctuations are marked by only one sample in the pollen core, so without additional sampling, random fluctuations due to changes in pollen deposition and preservation cannot be conclusively ruled out. Assuming background noise is not the primary explanation, three other explanations come to mind: i) treeline changes, ii) forest fire history, or iii) a combination of forest fire and insect attacks.

Windmill Lake is currently about 160 m below the *Picea* treeline. Because *Picea*, *Betula*, and *Alnus* are prolific pollen producers, a significant change in treeline is needed before a record is preserved in the Windmill Lake pollen core. Pollen data, because they reflect a regional signal, are not especially sensitive to treeline changes. Tenmile Lake (Figure 1.2), located in the Gulkana uplands near the Delta River, is located 5 km beyond treeline in *Betula* shrub tundra. However, (possibly because the lake is near lower elevation boreal forests), the pollen record consistently contains >30% *Picea* pollen from 9000 ^{14}C yr BP to the present (Anderson et al., 1994). This suggests that if the *Picea* fluctuations at Windmill Lake in fact reflect treeline changes, then significant and rapid treeline shifts must have occurred over a large area. Currently, this seems unlikely because there is no independent evidence (i.e., glacial advances and/or retreats) of transient large-scale climatic changes needed to initiate widespread treeline movement during the mid- to late Holocene.

A more likely answer to explain the fluctuations in the Windmill B core would be a mechanism that would rapidly change the relative proportions of *Picea*, *Betula*, and *Alnus*, but not necessarily affect the overall distribution of the boreal forest. One such mechanism would be repeated burning of the *Picea* forest. Today, forest fires play a significant role in determining the vegetation of the boreal forest (Viereck, 1983). The interval between *Picea* peaks is 400 to 500 years, that is 200-250 years between *Betula*/*Alnus* peaks and *Picea* peaks. 200 years is about the amount of time needed to reestablish *Picea* after a fire (Dyrness et al., 1986). For fire to drive the *Picea*/*Betula*/*Alnus*

cycles, they must burn primarily the *Picea*-covered areas, with relatively little burning occurring in the *Betula* and *Alnus*. While *Picea* forests do burn easily, because of the accumulation of fuels, as well as the high volatile content of the needles and sap, deciduous woodlands also burn, particularly in the spring, before green-up. Thus, if forest fires were a significant factor determining the pollen fluctuations at Windmill Lake, these fires would have needed to burn large areas of *Picea* in order to be seen in a lacustrine pollen record.

Pollen records from other lakes in central Alaska do not show the strong cycles seen at Windmill Lake, but that may be due partially to coarse sample resolution. Charcoal analysis has not been conducted at Windmill Lake, although superficial examination of the pollen slides indicates very low charcoal accumulation throughout the Holocene. This suggests that while forest fires may have played a role in the changing *Picea* abundance, other factors may have been equally important.

The rapid *Picea*/*Betula*/*Alder* cycles at Windmill Lake suggest short-term changes in the relative proportions of these species within the forest. If forest fire did not drive the cycles, perhaps a pathogen, such as repeated spruce bark beetle infestations played a role (or perhaps an interaction between beetle infestations with forest fires). Spruce beetle infestations often occur after forest fires because of the quantity of dead timber (Coulson and Witter, 1984). A combination of repeated forest fires and spruce beetle infestations may explain the fluctuating pollen signal at Windmill Lake. Additional corroborating evidence from other lakes and macrofossil evidence for either forest fire or beetle infestation are needed, however, to adequately test this hypothesis.

Stable carbon isotopes. The particulate organic matter in the Windmill B core probably originated from within the lake. The C/N ratios and nitrogen isotopes are somewhat ambiguous, but the texture of the sediments (very fine organic-rich mud) and the presence of abundant algae in some of the pollen preparations suggest the core is dominated by lacustrine organic matter. This indicates that the variations in carbon isotopes in the core mainly reflects

changes in the lake's water chemistry (see Appendix B for a general discussion of factors influencing $\delta^{13}\text{C}$).

The two primary factors affecting aquatic $\delta^{13}\text{C}$ are changes in the $\delta^{13}\text{C}$ of the DIC (dissolved inorganic carbon) entering the lake, and changes in the $\delta^{13}\text{C}$ of the DIC within the lake. At Windmill Lake, it is unlikely that much of the DIC entering the lake originated from the bedrock, as the watershed is underlain by a diabase intrusion. In addition, many Arctic lakes, (including Windmill during July 1996), are super-saturated in CO_2 relative to the atmosphere, suggesting that these lakes are degassing CO_2 to the atmosphere, instead of incorporating atmospheric CO_2 (Kling et al., 1991).

DIC can also enter lakes as rainwater on the lake surface, but mass balance models of modern water budgets from two central Alaskan lakes (Birch and Jan lakes) indicates that precipitation on the lake surface represents only 1/3 to 1/2 of the water inputs into the lakes, the remainder coming from watershed run-off and groundwater flux (B. Finney, pers. comm.). In addition, the $\delta^{13}\text{C}$ of lake water DIC from Smith Lake near Fairbanks, as well as a subalpine lake (Findley Lake) in the Cascade mountains in western Washington, are depleted in $\delta^{13}\text{C}$ relative to the atmosphere, also suggesting that atmospheric CO_2 is not a significant source for the lakes' DIC (Gu and Alexander, 1996; Rau, 1978).

Some of the $\text{CO}_{2\text{aq}}$ in Windmill Lake today is probably entering the lake through groundwater runoff. In non-carbonate regions, the $\delta^{13}\text{C}$ of the groundwater is influenced by the CO_2 respired from the vegetation and near-surface organic matter. C_3 plants, which dominate in northern latitudes, are significantly depleted in $\delta^{13}\text{C}$, with values ranging between about -30‰ and -22‰ (Bender, 1971), which explains the low $\delta^{13}\text{C}$ values in lake water DIC at Smith Lake as well as Findley Lake in the Cascade mountains.

The $\delta^{13}\text{C}$ signal of DIC is changed prior to sedimentation mainly by cycling through biological systems, which transform it into DOC (dissolved organic carbon) or POC (particulate organic carbon). The isotopic signal of aquatic particulate carbon varies mainly due to fluctuations of DIC within the lake water and photosynthetic rate. While the isotopic signal of lake water DIC can change due to external forcing, such as changes in groundwater

chemistry, or changes in the lake's water budget, the $\delta^{13}\text{C}$ of POC can also change due to growth rate and drawdown of $\text{CO}_{2\text{aq}}$ during periods of photosynthesis. As CO_2 becomes limiting, less fractionation takes place, and the resulting organic carbon is enriched in the heavy isotope. In addition, the CO_2 drawdown may initiate the transfer of atmospheric CO_2 into the lake, which also enriches the $\delta^{13}\text{C}$ of lake water DIC. Finally, in very productive lakes, bicarbonate may become an important component of the utilized carbon pool. All three of these factors result in elevated lake water $\delta^{13}\text{C}$, which is manifested in high $\delta^{13}\text{C}$ in DOC and POC (Gu et al., 1996).

To distinguish between isotopic changes due to variations in the input of respired CO_2 via the groundwater and runoff from those associated with productivity changes, carbon and nitrogen fluxes ($\text{g}/\text{cm}^2/\text{yr}$) were calculated to use as a proxy for past productivity. Flux measurements, however, are sensitive not only to changes in the concentration of the elements, but also to changes in bulk density and sedimentation rate. That is, increases in either bulk density or sedimentation rate will increase the flux, even if the C and N concentrations remain constant. As a result, the C and N concentrations and fluxes are used together as a proxy for productivity. However, the relative preservation of C and N can vary simply due to changes in sedimentation rate. Rapidly buried sediments, particularly those in low oxygen environments tend to have higher C and N than sediments in oxic or slowly buried settings (Calvert et al., 1991).

The base of Windmill B (before about 12,000 ^{14}C yr BP), is relatively heavy in $\delta^{13}\text{C}$, with values as high as about -24‰ (Figure 2.15). Productivity is low (as indicated by low C and N fluxes and concentrations) (Figure 2.14), suggesting that productivity changes were not driving the isotopic signal.

Today, groundwater $\delta^{13}\text{C}$ is probably most strongly affected by the dense vegetation cover of C3 plants and resultant soil development. Before 12,000 ^{14}C yr BP, however, the vegetation was significantly sparser, with abundant open ground and probably lower soil development; this may have resulted in more enriched $\delta^{13}\text{C}$ in the groundwater. In addition, changes in Windmill Lake's water budget may have altered the $\delta^{13}\text{C}$ of the lake water because of variations in the relative importance of groundwater, runoff, precipitation on

the lake, and evaporation. Reduced groundwater flux into the lake could have increased the $\delta^{13}\text{C}$ of the lake, because precipitation (which is enriched in $\delta^{13}\text{C}$) would have contributed proportionally more to the water budget.

Between 11,800 and 10,600 ^{14}C yr BP, $\delta^{13}\text{C}$ decreases by more than 6‰ (Figure 2.15), C and N fluxes increase, but the C/N ratios remains relatively low. This suggests a major shift in Windmill Lake's carbon cycle. The lowered $\delta^{13}\text{C}$ is probably due to higher groundwater flux of $\text{CO}_{2\text{aq}}$ into Windmill Lake. The vegetation at this time also shifts to warmer and more mesophytic species, also suggesting increased effective moisture.

A small increase in $\delta^{13}\text{C}$ (+1‰) at 10,400 ^{14}C yr BP may suggest a brief period of lowered groundwater flux, although C and N fluxes are high (Figures 2.14 and 2.15). The C and N fluxes are high mainly because of increases in bulk density, as well as slightly elevated C and N concentrations. Based on these considerations, the higher $\delta^{13}\text{C}$ is probably more the result of lowered water tables than higher productivity. The pollen of xerophytic species also increases at this time, indicating reduced effective moisture.

Between about 10,000 ^{14}C yr and 6000 ^{14}C yr BP, $\delta^{13}\text{C}$ returns to low levels, while C and N concentrations increase, C/N ratios increase, and C fluxes are stable. This suggests that lake productivity was probably increasing during this period, but that C was not limiting, so the $\delta^{13}\text{C}$ reflects mainly the $\delta^{13}\text{C}$ of the groundwater entering the lake.

By 6000 ^{14}C yr BP, however, significant shift in the lake's carbon cycle is marked by increased $\delta^{13}\text{C}$, increased C and N fluxes, and increased C/N ratios (Figures 2.14 and 2.15). These changes probably reflect higher productivity and more abundant macrophyte remains in the sediments. The higher quantity of macrophytes may reflect a combination of lower lake level during the latter half of the Holocene, in addition to infilling of the lake basin, which created more habitat for the macrophyte growth. The increased $\delta^{13}\text{C}$ probably reflects C limitation and $\text{CO}_{2\text{aq}}$ drawdown during photosynthesis.

Stable nitrogen isotopes. Theoretically, nitrogen isotopes should reflect changes in the $\delta^{15}\text{N}$ of the dissolved inorganic nitrogen (DIN) prior to entering the lake, as well as changes in the DIN and the production of organic nitrogen within the lake. However, most studies have focused on the $\delta^{15}\text{N}$

of nitrogen as it travels through the lacustrine N-cycle, with little attention paid to changes in the isotopic signal of N before entering the lake. One reason is probably because the isotopic changes of N within the lake are so large that they overwhelm any changes related to, for example, groundwater fluctuations or shifts in the water budget.

Nitrogen isotopes may delineate terrestrial and lacustrine N, with terrestrial autotrophs typically having low $\delta^{15}\text{N}$ values (mode, -1‰), and lacustrine autotrophs having slightly elevated $\delta^{15}\text{N}$ (mode, $+3\text{‰}$), although there is significant overlap between these two groups (France, 1995). Thus changes in $\delta^{15}\text{N}$, when examined together with C/N ratios, may reflect variability in the amount of terrestrial and aquatic plant remains preserved in a lake core.

Changes in sediment $\delta^{15}\text{N}$ can also reflect changes in the N residence time in the water column. $\delta^{15}\text{N}$ increases about 3‰ per trophic level (at least at the phytoplankton/zooplankton level) (Estep and Vigg, 1985; Gu et al., 1994). In addition, as N is repeatedly recycled through microorganisms, the $\delta^{15}\text{N}$ of dissolved inorganic nitrogen (DIN) increases because microorganisms preferentially excrete the heavier isotope, which is ingested by another organism, thus, with time (if terrestrial DIN is not abundant), the $\delta^{15}\text{N}$ of the lake's usable N pool increases.

As N becomes limiting, fractionation decreases (Peterson and Fry, 1987), so the $\delta^{15}\text{N}$ of the lacustrine autotrophs reflects the isotopic signal of their N sources (which maybe quite high, because of recycling).

Episodes of nitrification (the production of nitrate) and denitrification (the production of nitrite and nitrogen gas) (Horne and Goldman, 1994) strongly fractionate against ^{15}N (Peterson and Fry, 1987; Gu, 1993c), which, in lakes with low terrestrial DIN input, increases the $\delta^{15}\text{N}$ of both the N source and the product.

Finally, denitrification can also reduce the $\delta^{15}\text{N}$ of the lake because it removes usable N from the lake's N pool. With a continual supply of watershed-derived DIN, relatively more of the lake's DIN will have the lower $\delta^{15}\text{N}$ signal of terrestrial N, instead of the higher $\delta^{15}\text{N}$ of recycled N. It is still not clear, however, how much terrestrial DIN enters subarctic lakes today.

The low $\delta^{13}\text{C}$ of modern lake sediments and plankton suggests significant amounts of terrestrial DIC is present, implying that terrestrial DIN is also present. Lakes can survive by recycling N during periods of intense photosynthetic activity (Gu and Alexander, 1993b), but ultimately, usable N is lost to the biological system, and must be replenished from terrestrial or marine sources.

From about 13,000 ^{14}C yr BP, to about 10,800 ^{14}C yr BP $\delta^{15}\text{N}$ values are close to the median for the whole core, varying between about 1.3 and 1.5‰ (Figure 2.15). A single sample at about 12,500 ^{14}C yr BP has very little $\delta^{15}\text{N}$ (about 0.5‰) and may be an analytical error because of low N concentrations. C/N ratios are also low for this period, indicating that algae are the main N source in the sediment.

$\delta^{15}\text{N}$ achieves its highest values for the entire core between about 10,600 and 9800 ^{14}C yr BP. This coincides with high *Pediastrum* concentrations and rising C and N concentrations (Figures 2.8, 2.14, and 2.15). Some of the increased $\delta^{15}\text{N}$ might be due to the high quantities of *Pediastrum* (green algae typically have higher $\delta^{15}\text{N}$ values than blue green algae [Gu and Alexander, 1993a]) but this relationship should not be overemphasized because it is impossible to know how large a role the *Pediastrum* played within the entire algal community.

The increased C and N concentrations suggest the lake was more productive than earlier periods. However, the isotopes and pollen data suggest this was a period of relative aridity and the flux of terrestrial N into the lake was reduced. Thus, the high $\delta^{15}\text{N}$ probably reflects the additional recycling of N, which elevated the $\delta^{15}\text{N}$ of the *Pediastrum* and other algae.

From about 9600 to about 6000 ^{14}C yr BP, $\delta^{15}\text{N}$ decreases while C and N concentrations, as well as C/N ratios (particularly at the end of the period), increase (Figure 2.14 and 2.15). Productivity probably reached modern levels at this time. The lowered $\delta^{15}\text{N}$ suggests reduced recycling of N, possibly a result of increased terrestrial DIN input, as is suggested by the low $\delta^{13}\text{C}$ and higher C/N ratios. Alternatively, the $\delta^{15}\text{N}$ may reflect a change in the algal community, where blue-green algae become an important part of the ecosystem.

After about 6000 ^{14}C yr BP, $\delta^{15}\text{N}$ gradually increases to modern levels. C and N concentrations decrease slightly about 4000 ^{14}C yr BP, then increase to modern levels. C/N ratios are highest for the entire core between 5600 and 2500 ^{14}C yr BP, then decrease to modern levels (Figures 2.14 and 2.15). The slightly higher $\delta^{15}\text{N}$ may reflect an increase in nitrification and recycling of N in the lake, despite the apparently higher flux of terrestrial material. It is probable that the lake was highly productive at this time, and that nitrogen use was not offset by increased terrestrial N.

Sediment $\delta^{15}\text{N}$ may be affected by watershed vegetation and soil processes. However, the overall $\delta^{15}\text{N}$ values in the core are somewhat higher than average terrestrial values, suggesting that any N reaching the lake is transformed prior to sedimentation. In addition, there is no change in either total N or $\delta^{15}\text{N}$ when *Alnus* (a symbiotic N-fixer) appears on the landscape. Soil nitrogen around N-fixers increases due to litter-fall (Binkley et al., 1985; Virginia and Delwiche, 1982). However, at Windmill, total nitrogen was steadily increasing before the arrival of *Alnus*, and does not show any change with the appearance of *Alnus*. The $\delta^{15}\text{N}$ of N-fixers tends to be lighter than non-N-fixers, however, where $\delta^{15}\text{N}$ has been measured in the soil around the N-fixer, there is often no significant change in the soil $\delta^{15}\text{N}$ (Binkley et al., 1985), indicating other factors effect the soil nitrogen, and that an increase of a nitrogen-fixer on the landscape is unlikely to effect the $\delta^{15}\text{N}$ signal within the lake.

Lake level changes. Lake levels are recorded by lithologic changes in the cores, as well as by the kinds of organic material that are preserved in them. Shoreline sediments are sandy, often with abundant reworked terrestrial and macrophyte plant remains (Digerfeldt, 1986). However, in lakes with a sedge mat ringing the edge, shoreline sediments can also be organic-rich with abundant macrophyte remains. In addition, the macrofossils from emergent aquatic vegetation may also provide accurate lake level information, because these plants grow at specific water depths, and in most cases, their seeds or macrofossil fragments do not travel far from the parent plant (Harrison and Digerfeldt, 1993; Birks, 1973)

In deep water, the sediments are dominated by finely divided organic detritus (gyttja), which is only deposited where wave turbulence does not reach to the sediment. The transition from sandy shoreline sediments to gyttja (the "sediment limit") occurs at a range of water depths, depending mainly on lake size and shoreline morphology. Areas with a long fetch will have a deeper sediment limit than areas with a shorter fetch (Digerfeldt, 1986).

The presence of pebbles and sand in the lowest sediments of the Windmill B core suggests the sediments are either a remnant of the late Wisconsinan moraine, or they are reworked morainal sediments which were deposited in the basin. In either case, by about 13,000 ^{14}C yr BP, the presence of *Pediastrum* in these deposits suggests the lake basin contained some water. The mixture of inorganic silt, sand, and pebbles, as well as rare reworked macrofossils indicate that the lake was very shallow and unproductive at this time. The coring location was probably not far from the lake edge, suggesting about an 8 m drop in water level (Figure 2.16).

By about 11,700 ^{14}C yr BP, increasing gyttja in the central core suggest the lake level had risen somewhat. $\delta^{13}\text{C}$ values also decrease at this time, suggesting higher ground water flux into the lake. Core D was still subaerial at this time, indicating the lake was more than 4 m below modern, possibly as much as 7 m below modern (Figures 2.3 and 2.16).

In core D, a radiocarbon date ($10,980 \pm 50$) on seeds from shallow water or shoreline taxa (*Carex*, *Potamogeton*, and *Ranunculus*) suggest the lake edge was nearby, indicating a lake level 4 to 5 m below modern (Figure 2.3). The lake probably remained at this level for a number of years, because silty gyttja and lenses of sand (indicating shallow water deposits) extend about 30 cm above the dated seeds.

In Windmill D, an undated transition to humified peat (with abundant moss fragments) occurs about 30 cm above the dated seeds. The humified peat extends for about 150 cm, above which a sedge peat is preserved. The peats suggest increasing water level and/or shoreline stabilization.

Above the peat lies a 10 cm-thick gyttja deposit with macrofossils. The transition to the gyttja is dated about 6000 ^{14}C yr BP. The gyttja suggests the

Windmill Lake

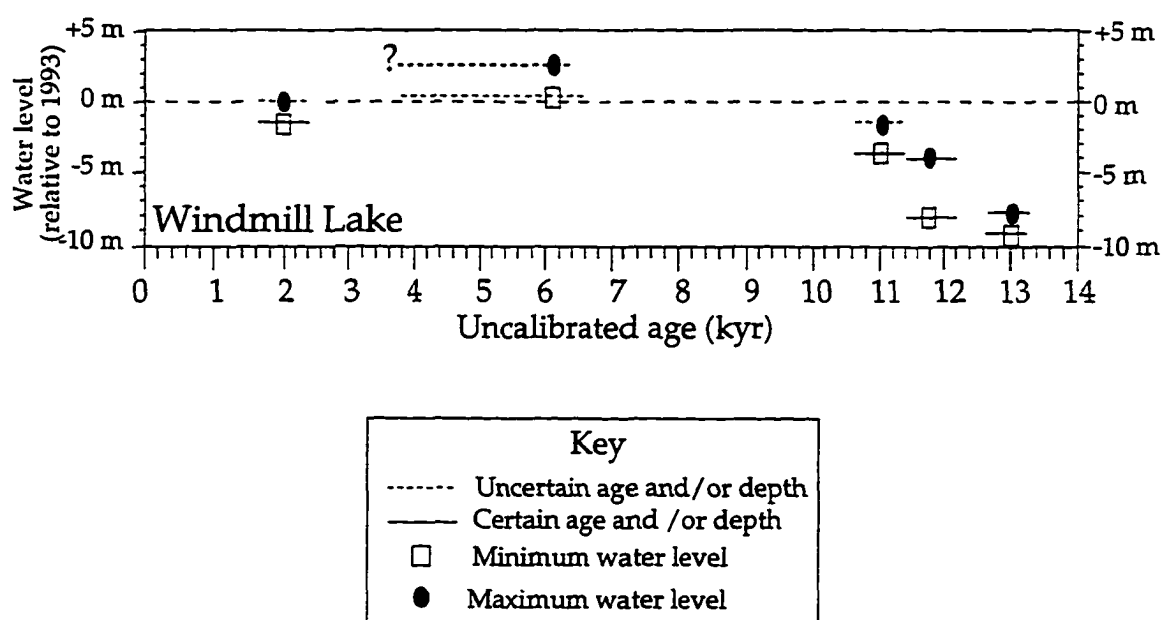


Figure 2.16: Windmill Lake lake level changes.

water level was higher than today. Afterwards, the relative amount of macrofossils gradually increases, until about 60 cm from the top of the core, where the sediments become peaty, extending to the core top. These changes indicate a gradual decrease in water level.

The water level after 6000 ^{14}C yr BP was probably only slightly above modern. A morainal deposit dams the lake at its outlet, the top of the dam is not more than 1-2 m above the modern water level. In addition, a probable shoreline is preserved on the northeast side of the lake, which also lies about 1-2 above the modern shore.

In Windmill C, much of the core is of gyttja mixed with macrofossils. The core does not appear to show a period of higher water level (a reduction in macrofossils), but that may be because the core is in a comparatively insensitive location. In the lake bottom today, the zone where gyttja and macrofossils (especially moss fragments) are deposited together is fairly wide (moss today extends between about 10 cm and 170 cm water depth), thus a core located in the middle of the zone may not record a 1 m increase in water level.

Summary

In sum, the multiple proxy data from Windmill Lake provide a detailed record of vegetation and climate change from about 14,000 ^{14}C yr BP to the present.

From before 13,000 to about 11,700 ^{14}C yr BP, pollen, isotopic, and stratigraphic data indicate the lake was at a low stand, as much as 8 m below modern levels. The pollen data suggest the landscape was sparsely covered by herbaceous taxa with xeric preferences.

At about 11,700 ^{14}C yr BP, the ecosystem of the lake and its watershed changes dramatically. Carbon isotope data indicate the lake was receiving more ground water, although the stratigraphy of the side cores suggest that lake level was still more than 4 m below modern. The pollen data also indicate increasing moisture and warmth with the abrupt appearance of *Betula* in the record. Birch may not have been very widespread on the

landscape; herbaceous taxa continued to occupy xeric sites, while the birch occupied the more mesic sites.

Between about 10,800 and 9800 ^{14}C yr BP a short but significant change in the dataset occurs. The carbon isotopes and pollen data suggest a brief interval of increased aridity. The timing of these changes suggest they occurred during the Younger Dryas chronozone. The pollen data indicate drier conditions or cooler temperatures (increases of herbaceous taxa at the expense of shrubby taxa). Lake level was probably about 4 m below modern

The *Populus* subzone between about 9500 and 8500 ^{14}C yr BP marks the first arrival of trees in the Windmill Lake watershed. *Populus*, a thermophilous species compared to the extant *Betula*, was probably reacting to increasing summer warmth, in addition to local factors, such as seed source, open ground, and competition from other taxa.

The appearance of *Picea* in the watershed about 8300 ^{14}C yr BP marks the beginning of a significant change in the vegetation and landscape ecosystem. *Alnus* probably wasn't present until about 6800 ^{14}C yr BP, and *Picea mariana* probably wasn't significant until about 6000 ^{14}C yr BP. A high lake level after 6000 ^{14}C yr BP, suggests greater than modern effective moisture at this time.

At 3800 ^{14}C yr BP, the pollen data indicates *Picea* was very common and more abundant than today. The landscape from 5000 to 3800 ^{14}C yr BP was probably covered by a dense, spruce-rich boreal forest. However, the decrease in spruce concentrations after 4000 ^{14}C yr BP, combined with other widespread proxy data for climatic cooling, may indicate that the paludification and permafrost development we see today, started to form after 4000 ^{14}C yr BP.

Finally, the abrupt fluctuations in the *Picea*, *Betula*, and *Alder* indicate other factors were affecting the boreal forest. It is unlikely treeline changes caused the fluctuations, which means internal dynamics within the boreal forest are the likely causes. Two possible mechanisms are forest fire and/or spruce beetle infestations. Both mechanisms would heavily affect the spruce, and at Windmill Lake, the interval between low spruce to high spruce frequencies is roughly similar to the time required today to regenerate a spruce forest after a forest fire.

CHAPTER 3--DUNE LAKE

Introduction

Dune Lake, because it is located within a stabilized dune field, provides a sensitive record of effective moisture. Sand dunes are very well drained, and plants can be moisture-limited on these sites, so that small changes in effective moisture may have a strong effect on the local vegetation, which can be seen in the pollen records.

The water level in Dune Lake is controlled by a combination of regional groundwater, as well as local changes in precipitation and surface run-off. Thus, by comparing the lake level changes at Dune Lake with other lakes in central Alaska, regional and local differences in effective moisture can be identified.

Study Site

Dune Lake (64° 25'N, 149°54'W) is located in the Tanana River alluvial plain approximately 44 km southwest of Nenana (Figure 2.1). It is about 12 ha in area, with a maximum depth of about 9 m (Figure 3.1)

The lake lies at an elevation of about 134 m, within an extensive late Quaternary dune field. The tallest dune crests are approximately 60 m above the current water level. The dunes surrounding Dune lake and other nearby areas indicate the dominant wind direction during dune formation was from the northeast (Collins, 1985; Lea, 1996). The distribution of the sand dunes (adjacent to rivers draining the north slope of the Alaska Range) suggests these rivers were the primary sediment source. The timing of dune stabilization is poorly known, although basal radiocarbon dates from Dune Lake suggest that significant dune migration had probably ceased before 10,000 ¹⁴C yr BP (to allow lake formation in the interdunal basin). The dunes probably were not completely stabilized until the Holocene. This agrees well with sand dune and sand sheet stabilization dates in eastern Alaska and on the North Slope (Fernald, 1965; Carter and Galloway, 1984; Carter, 1983).

The modern vegetation at the lake is a mixture of open woodland/scrub and closed boreal forest. A forest fire in 1981 burned the southern end of the

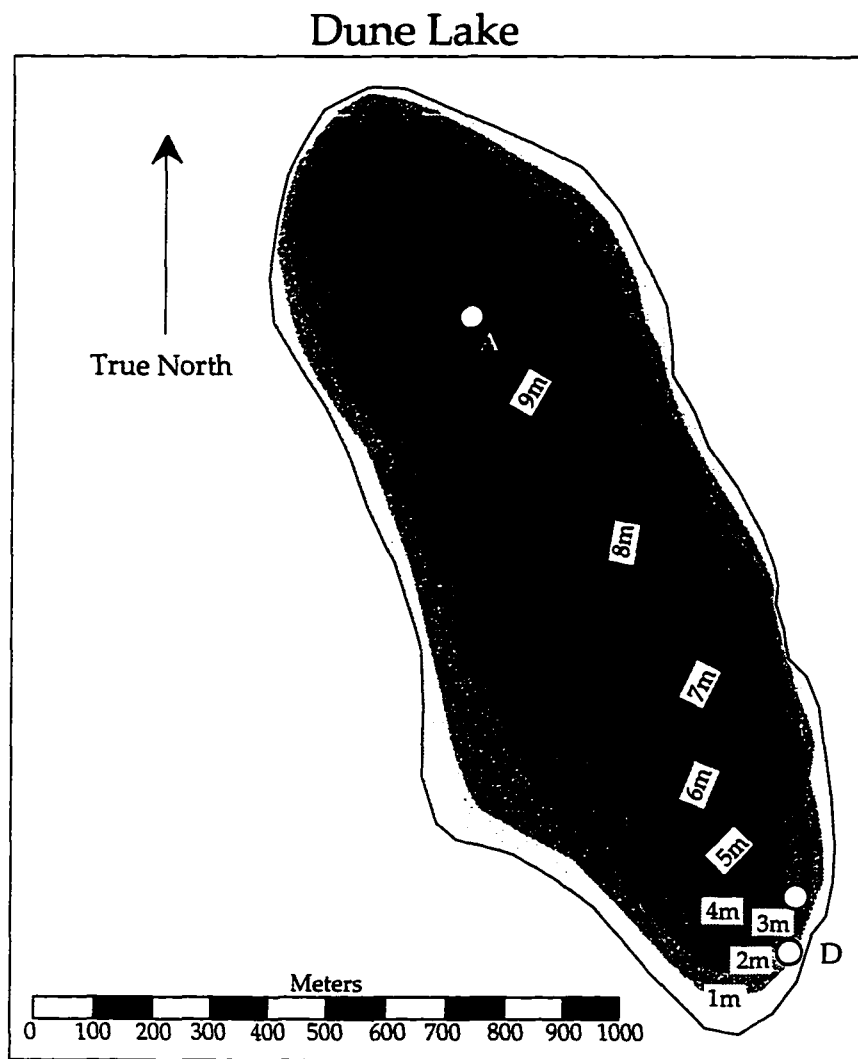


Figure 3.1: Dune Lake bathymetry and core locations.

lake; this is where the scrub (*Salix* spp., *P. tremuloides*, and *B. Papyrifera* saplings <2m tall) grows today. The ground cover is dominated by *Arctostaphylos uva-ursi* (bear berry), *Solidago* spp. (golden rod), Poaceae, *Bupleurum triradium*, and mosses. The open woodland (*P. tremuloides*, *B. papyrifera*, *P. glauca* trees, along with *Salix* spp., and *Shepherdia* bushes) grows on the well-drained beach ridges, south-facing dune slopes and dune crests. *A. uva-ursi*, *Geocaulon lividum*, *Vaccinium vitis-idea* (low bush cranberry), and grasses dominate the ground cover in these locations. The vegetation on some south-facing dune slopes is discontinuous, with common *Elaeagnus commutata* (silver berry) bushes.

The closed boreal forest is found in the less well-drained areas and on the north-facing dune slopes. Areas with moderate drainage have a mixed forest of *P. glauca*, *B. papyrifera*, and occasional *P. mariana*, along with a shrub layer dominated by *Salix* and *Rosa*. The ground cover is a mixture of *V. vitis-idea*, *Cornus canadensis* (bunch berry), and mosses. Poorly drained areas are dominated by *P. mariana* with a mossy ground cover.

Because the lake water level at Dune Lake is currently rising, an extensive *Carex* mat has not developed on the shoreline. Instead, the shoreline is dominated by drowned trees (*Picea* and *Betula*) with a shrubby vegetation of *Salix* spp, *Vaccinium uliginosum* (blueberry), *Ledum groenlandicum* (Labrador tea) and *V. vitis-idea*. Other taxa include *Equisetum*, grasses, and infrequent sedges.

Emergent vegetation is common only in the areas with a gently sloping shore line (west side, as well as north and south ends of the lake). Where present, the aquatic vegetation is dominated by *Polygonum amphibium* and *Carex* spp.; *Scirpus* spp. and *Sparganium* spp. are present in low quantities. At the south end of the lake, a well developed (floating?) mat of *Typha latifolia* and *Carex* spp. is also present.

Submerged vegetation (<2m water) includes *Potamogeton* spp. and *Myriophyllum* spp. *Ceratophyllum* probably also grows at Dune Lake, as its distinctive leaf-edge spines were recovered from the core tops.

Many of the submerged aquatics have carbonate crusts on their leaves. However, lake water chemistry indicates only the top 5 to 6 m of the water column has a basic pH, suggesting that the carbonates produced during

photosynthesis are only deposited at shallow depths. Measurements of dissolved oxygen taken during the summer thermocline show that only the bottom 1-2 m of the water column are anaerobic, and that during the fall and spring turnovers, the lake is completely mixed.

A series of raised beaches surround the lake, indicating past episodes of high water; the highest beach is approximately 10 m above the current (1994) lake level. In addition, several submerged beaches have also been recognized, the lowest of these is about 6 m below the modern lake level (Finney et al., 1995).

Finally, interviews with local residents indicate the water level has risen about 3 m since the early 1970s. This may be partly explained by higher than average summer precipitation since that time, but also by the increased run-off into the lake after the 1981 forest fire (Barber and Finney, 1995).

Results

A series of three Livingstone cores were collected from Dune Lake (Figures 3.1 and 3.2). Pollen analysis was conducted on Dune A and Dune C, while preliminary macrofossil analysis was also conducted on Dune C. Dune A was collected in April of 1993, in 890 cm of water, with a core length of 621 cm. Dune C was collected in August of 1994 in 266 cm of water, with a total length of 538 cm.

Core sediment descriptions. Dune A is the longest of the cores and provides a detailed sedimentary history of Dune Lake (Figure 3.3). From the core top (890 cm) to about 1375 cm (all measurements from the water surface), the sediments are a dark brown to tan gyttja. Laminations can be seen throughout much of the gyttja, although they are strongest 1129-1147 cm, and 1177-1355 cm. The laminations consist of alternating light and dark layers; smear slides indicate the light layers are composed mainly of diatoms.

At 1375 cm, thin layers of sand interrupt the gyttja. At 1385 cm the sediments are mostly sand, with interspersed layers of silt and sandy silt (Figure 3.4). Below 1460 cm the sediments are completely sand, which becomes coarser in the bottom 30 cm of the core.

Two tephras are present in Dune A, at 1248 cm and at 1314 cm. The tephra at 1248 cm (<1 mm thick) may be the Jarvis Creek Ash, but because the layer is

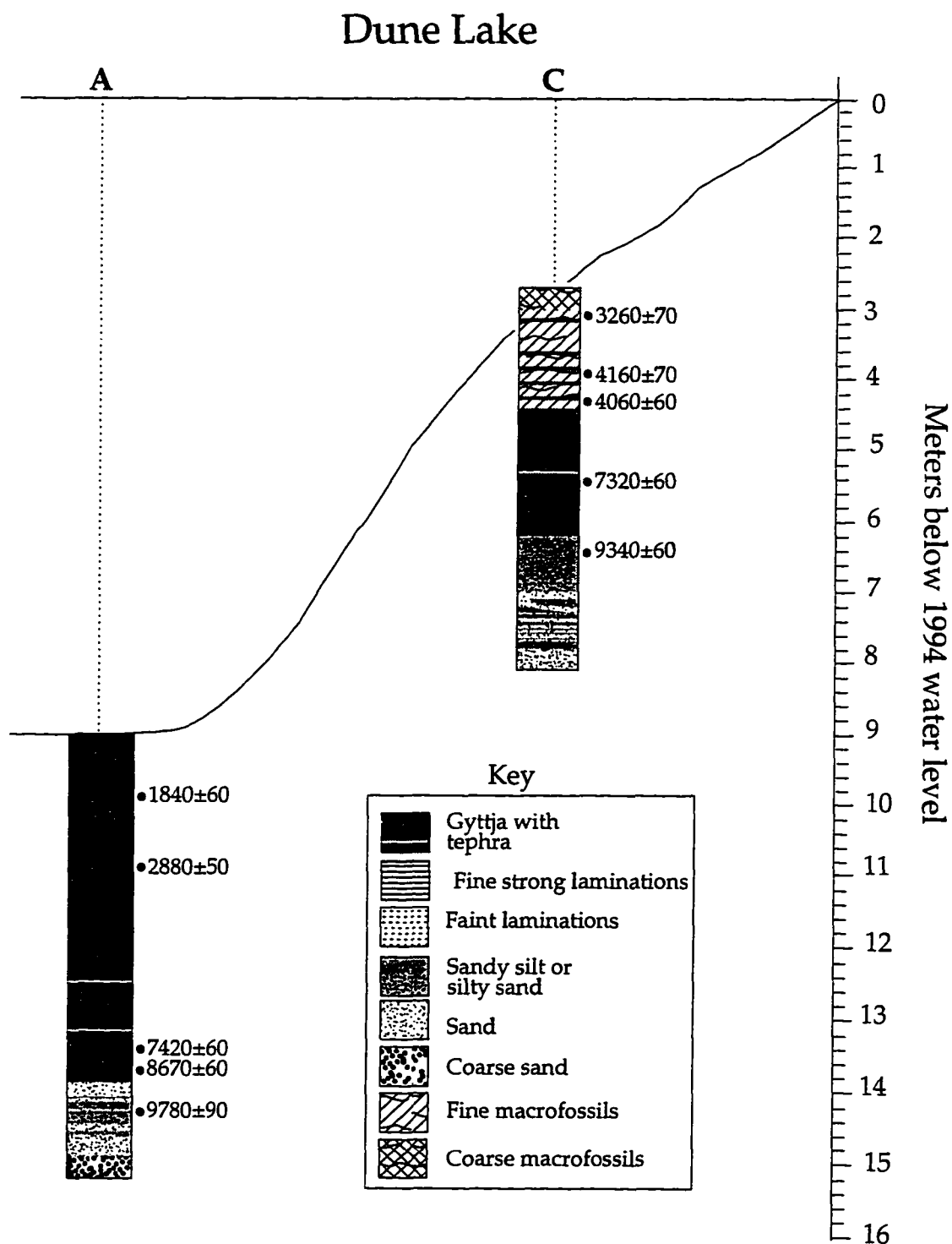


Figure 3.2: Dune Lake core stratigraphies with uncalibrated radiocarbon dates.

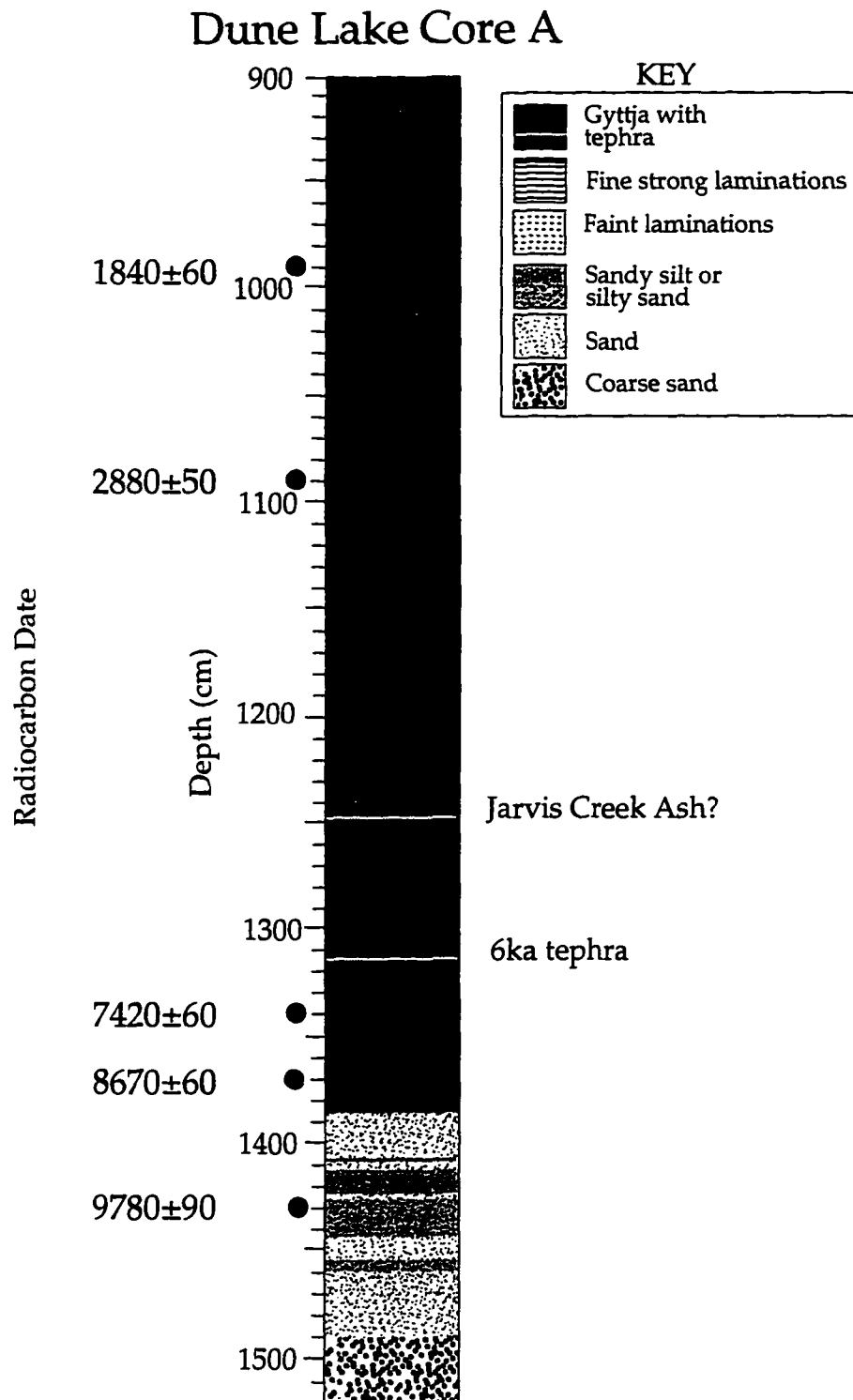


Figure 3.3: Dune A core stratigraphy and uncalibrated radiocarbon dates.

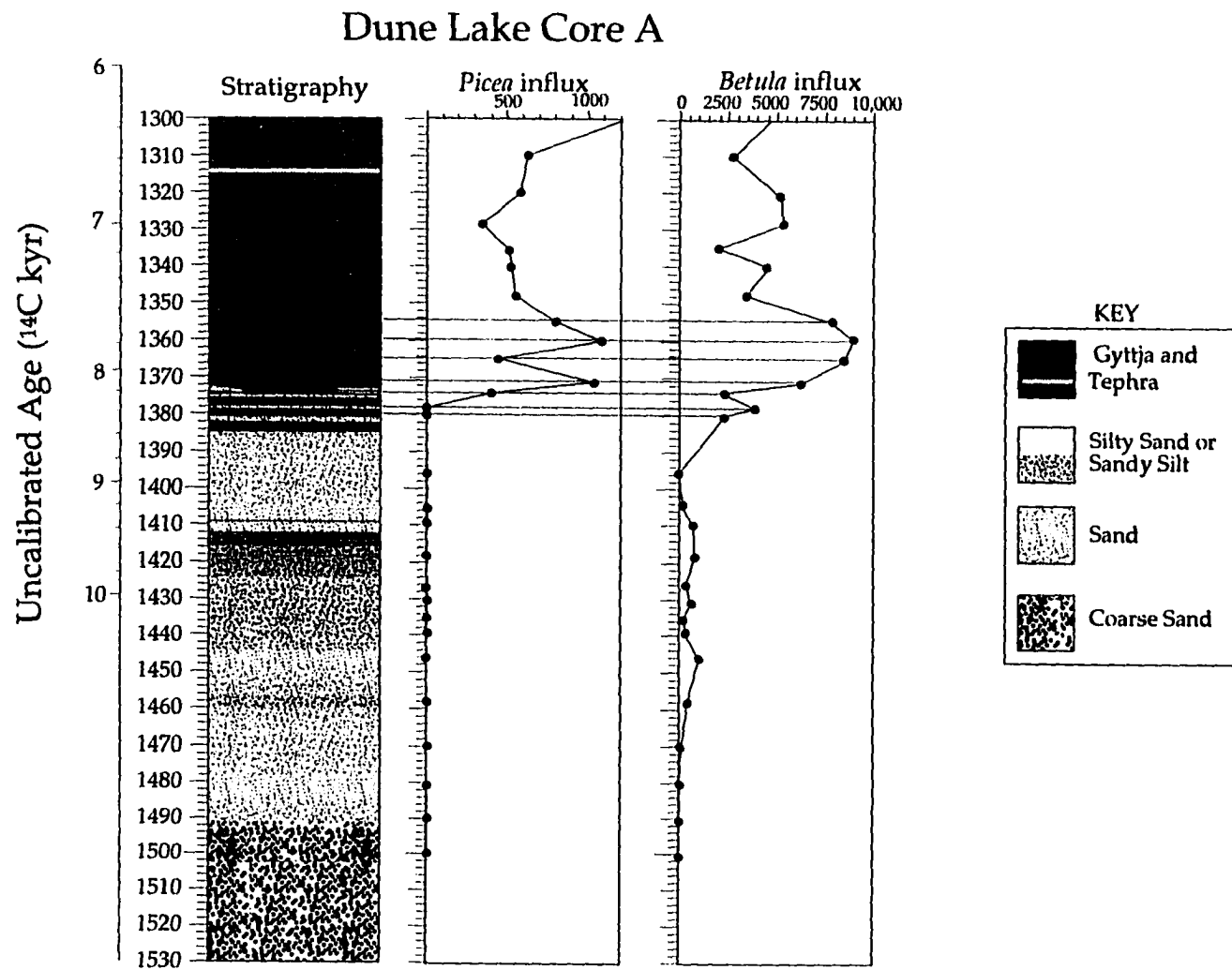


Figure 3.4: Dune A, detailed stratigraphy and influx of *Picea* and *Betula*.

very thin and somewhat diffuse, it has not been identified with microprobe analysis. The tephra at 1314 cm (2 mm thick) has been identified by microprobe analysis as a mid-Holocene tephra that is also preserved in Denali National Park (Child, n.d.). This tephra was initially discovered at Wonder Lake and a nearby pond; at these localities it is closely dated by AMS techniques to about 6000 ^{14}C yr BP. For the purposes of this paper, this tephra is informally named the 6ka tephra.

Dune C, taken in 266 cm of water, provides a sensitive record of past lake level changes (Figure 3.5). The coring location was subaerial as recently the early 1970s, when the water level was about 3m below the 1994 level.

From the core top to about 300 cm, the sediments consist of coarse fibrous plant remains, probably *Carex*. From 300 cm to 440 cm, the sediments continue to be fibrous, but the fragments are significantly smaller and are interspersed occasionally with levels of gyttja mixed with macrofossils (Figure 3.5). From 440 cm to about 620 cm the sediments are predominantly gyttja with scattered macrofossils throughout. At 618 to 619 cm a lens of fine sand interrupts the gyttja. Between 620 and 685 cm, the sediment is silty gyttja with layers of sand. At 685 cm the sediments grade into organic-poor sandy silt, and then into sand at about 700 cm.

From 700 cm about 782 cm the sediments are sandy, with interspersed layers of silt and sandy silt. The core stratigraphy between 732 and 772 cm is quite complex, with alternating layers of fine sand, coarse sand and sandy silt. A sandy fining upward sequence is also preserved between 764 cm and 761 cm. This complex stratigraphy may suggest multiple sediment sources. The fining upward sequence and the rapid textural changes suggest that much of this part of the core may be a fan deposit or reworked sediment from the shoreline. In addition, the sand layers may also reflect episodes of eolian deposition.

From 782 cm to the core base, the core is composed of inorganic massive medium sand, probably unmodified dune sand.

One tephra at 530 cm was recognized in this core, its stratigraphic depth in relation to adjacent radiocarbon dates suggests it is the 6ka tephra. The Jarvis Creek Ash was not recognized, probably because it was reworked by wave action along the Dune Lake shoreline.

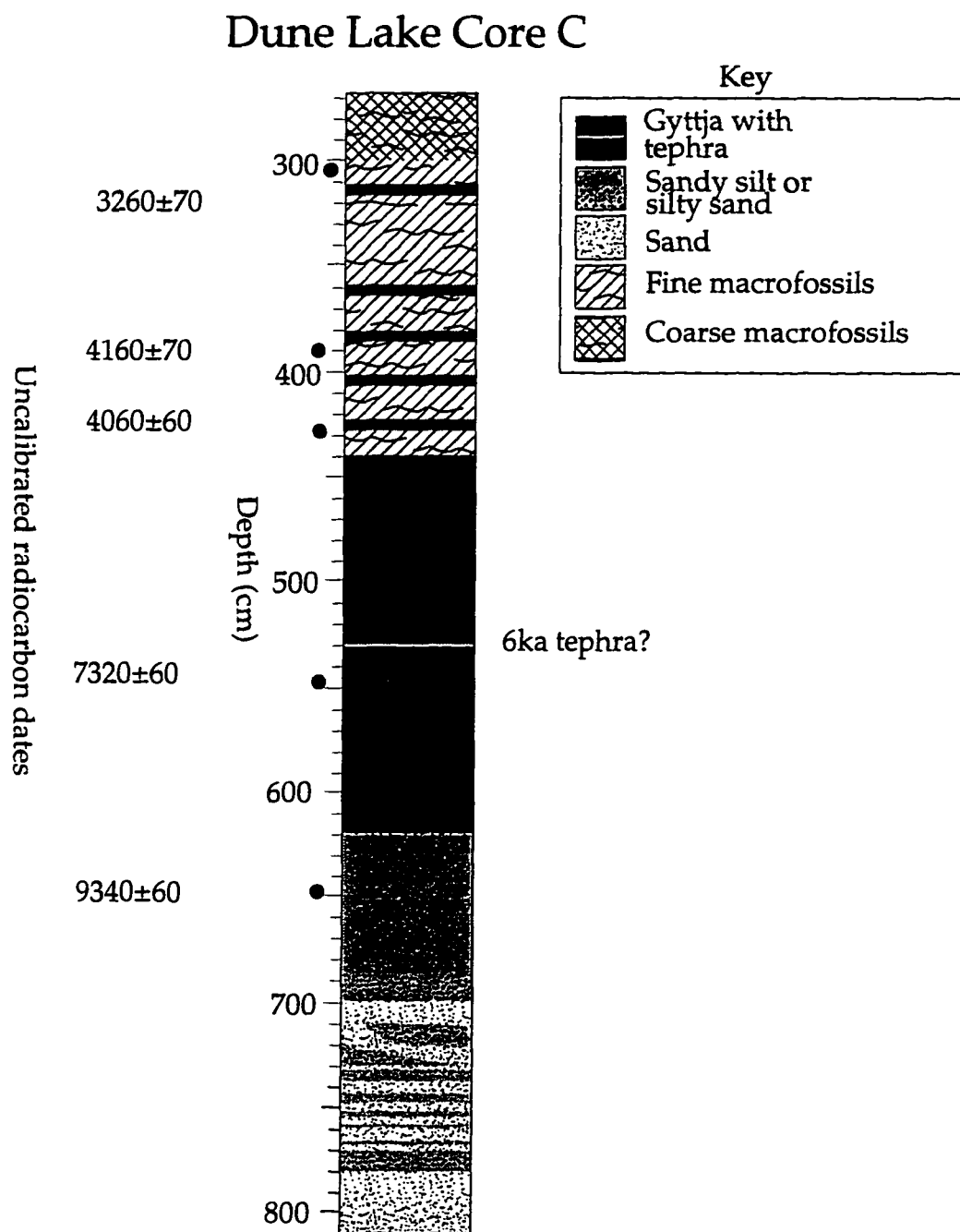


Figure 3.5: Dune C core stratigraphy and uncalibrated radiocarbon dates.

Radiocarbon dates and age models. A total of 10 AMS dates were processed from Dune A and Dune C (Tables 3.2 and 3.4), in addition to two dates on modern aquatic plant remains (Table 3.1).

Table 3.1: Radiocarbon dates on modern aquatics from Dune Lake.

Material dated	%Modern C	Lab No.	Comments
Live aquatic vegetation	102.9±0.8	CAMS-17018	aquatic carbon
Plankton tow (phytoplank. + zooplank.)	103.7±0.8	CAMS-17019	aquatic carbon

The dates on the modern aquatic plant remains indicates Dune Lake has a significant hard water or reservoir effect (Abbott and Stafford, 1996). The ^{14}C activity of the 1994 atmosphere (when the samples were collected) was 113% modern (due to additions of bomb-generated ^{14}C into the atmosphere). The reduced ^{14}C activity in the Dune Lake organics indicates the lake's dissolved inorganic carbon pool (DIC) is currently not in equilibrium with the atmosphere, and that radiocarbon dates on organisms that use lake water CO_2 may be too old. For this reason, terrestrial carbon sources were used for AMS dating. At Dune A (Table 3.2), most of the dates are on concentrated *Picea* pollen, using the method described in Appendix A. Only the oldest date is on terrestrial woody remains (probably bark).

Table 3.2: Radiocarbon dates from Dune lake, core A.

Depth (cm)	Material dated	mgC	Date	Lab No.	Comments
987-993	<i>Picea</i> pollen	0.98	1840±60	CAMS-29561	terrestrial carbon
1088-1091	<i>Picea</i> pollen	1.08	2880±60	CAMS-29562	terrestrial carbon
1336-1339	<i>Picea</i> pollen	1.12	7420±60	CAMS-29559	terrestrial carbon
1362.5-1368.5	<i>Picea</i> pollen	1.24	8670±60	CAMS-22020	terrestrial carbon
1427-1429	bark	0.17	9780±90	CAMS-31082	terrestrial carbon

The Dune A radiocarbon dates were calibrated using CALIB Rev 3.0 by Stuiver and Reimer, 1993 (Table 3.3). The calibration increases the ages of the dates, particularly with the oldest radiocarbon determinations.

Table 3.3: Dune A calibrated ages.

Uncalibrated date	Intercept BP	1 S.D. calibrated range (BP)
1840±60	1740	1701-1832 ^a
2880±50	2971	2891-3069 ^a
7420±60	8160	8126-8306 ^a
8670±60	9560	9522-9816 ^a
9780±90	10,980	10,943-11,001 [†]

^aCalibrated with Stuiver and Pearson (1993) bidecadal bristle cone pine calibration .

^aCalibrated with Pearson et al., 1993 and Linick et al., 1986, bristle cone pine and German and Irish oak calibration.

[†] Calibrated with Kromer and Becker (1993) German oak and pine calibration.

Five AMS dates were also obtained for Dune C (Table 3.4). The plant fragments were sieved from 2 cm-thick core slices using a 500 μ screen. The samples were a mixture of aquatic and emergent plant remains that had incorporated both terrestrial and aquatic carbon. A reversal in the radiocarbon dates at about 4000 ^{14}C yr BP may reflect not only the presence of old carbon within the lake, but also a change in the relative proportion of terrestrial and aquatic carbon in the samples.

Table 3.4: Radiocarbon dates from Dune lake, core C.

Depth (cm)	Material dated	mgC	Date	Lab No.	Comments
305	plant frags $\geq 500\mu$		3250±70	CAMS-17020	terr. + aquat. carbon
392	plant frags $\geq 500\mu$		4160±70	CAMS-17021	terr. + aquat. carbon
429	plant frags $\geq 500\mu$		4060±60	CAMS-17022	terr. + aquat. carbon
547	<i>Picea</i> cone		7320±60	CAMS-17023	terrestrial carbon
648	wood		9340±60	CAMS-17024	terrestrial carbon

An age model was constructed only for Dune A (Figure 3.6). Dune C does not have an age model because of the reversed radiocarbon dates in the middle Holocene.

With Dune A, a ^{210}Pb date at the base of one of the surface cores (140 yr BP) has been spliced on to the top of the core to provide an upper limiting date. The age for the 6ka tephra has been included in the age model, although it was not independently dated at Dune Lake. Comparison of the various models for Dune A (interpolated, linear, 2nd order polynomial, 3rd order polynomial) indicates that 3rd order polynomial ($r^2 = .989$) has the best fit with the radiocarbon dates (Figure 3.6).

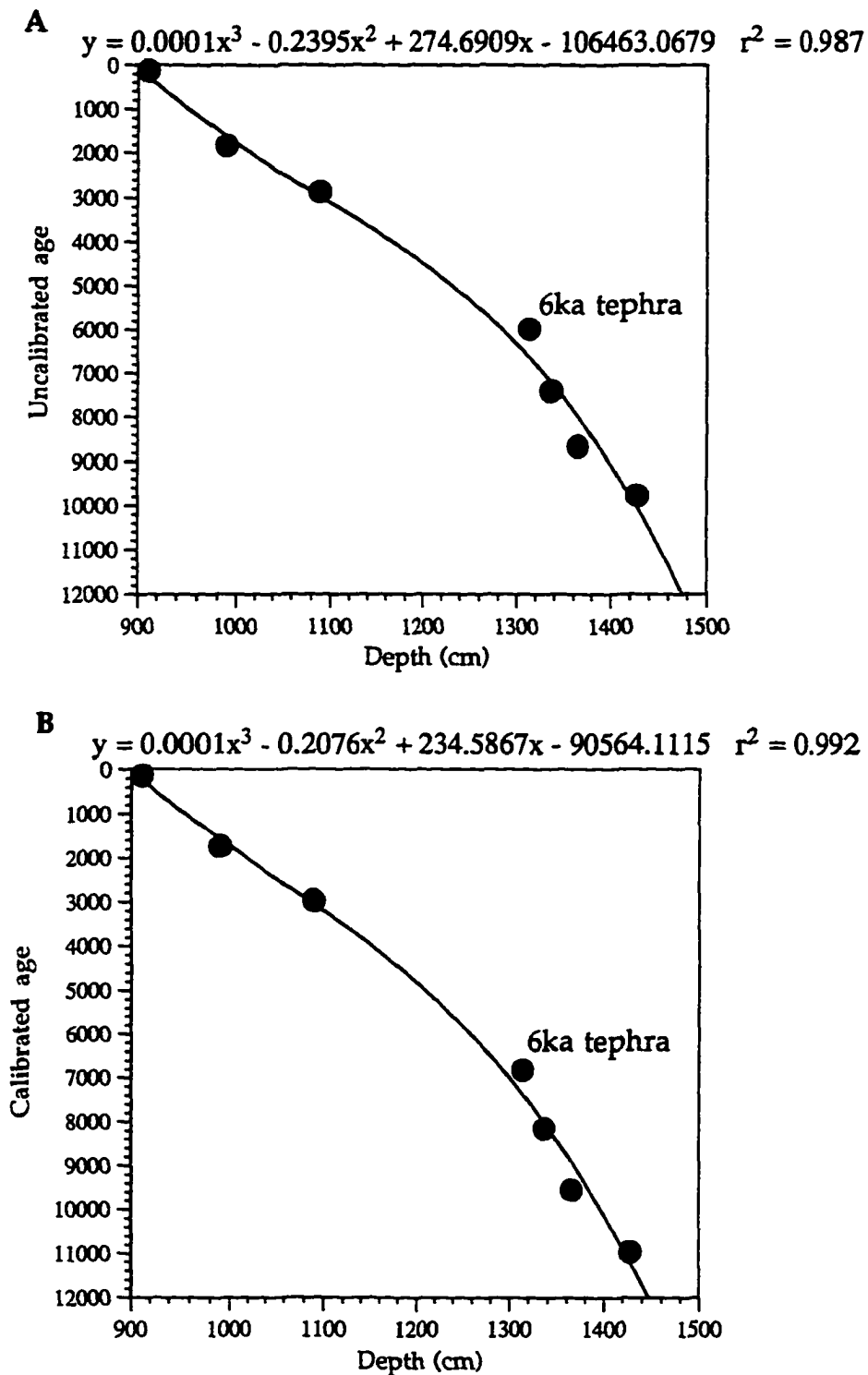


Figure 3.6: Uncalibrated (A) and calibrated (B) age models for Dune Lake core A.

Loss-on-ignition (LOI) and magnetic susceptibility. LOI and magnetic susceptibility reflect the varying organic content of the cores. In all the cores, magnetic susceptibility decreases at the core base. This does not reflect increased organic material, rather, the sandy sediments had slumped in the core holders, resulting in anomalously low magnetic susceptibility readings. The LOI and magnetic susceptibility analyses were conducted by Bruce Finney and Valerie Barber, and are used here with their permission.

At Dune A, the LOI (500°C) from the core base (>10,000 ^{14}C yr BP) to about 8300 ^{14}C yr BP is about 0%, while from about 8300 ^{14}C yr BP to about 3100 ^{14}C yr BP, LOI is about 30%. LOI then increases to 40%-50%, extending to the core top (Figure 3.7).

Magnetic susceptibility (MS) in Dune A is high at the core base, but rapidly decreases between 8500 and 8300 ^{14}C yr BP. Subsequently, the MS continues to gradually decrease to the core top. The gradual decrease of MS during the Holocene indicates that less and less inorganic silt and sand was being deposited at Dune Lake. The MS data also clearly mark the 6ka tephra, as well as the possible Jarvis Creek Ash.

LOI (850°C) provides an estimate of carbonate content (Dean, 1974; Bengtsson and Enell, 1985). LOI (850°C) is low at the core base, but increases gradually to about 5% by about 5000 ^{14}C yr BP (Figure 3.7). Subsequently the LOI (850°C) decreases slightly, but by about 2500 ^{14}C yr BP, it increases to 5%-6%, subsequently fluctuating markedly between 6% and 2%. Bivalve shells were not recognized in Dune A, indicating that these fluctuations in LOI may reflect bicarbonate precipitation during active photosynthesis.

At Dune C, LOI and MS, provide additional insights into the core stratigraphy (Figure 3.8). At the core base, LOI (500°C) is close to 0%, but gradually increases to 5% by about 660 cm (approx. 9300 ^{14}C yr BP). LOI (500°C) then drops to near 0%, but rapidly increases to 15% between 640 and 660 cm. LOI (500°C) remains about 15% until about 540 cm, then briefly decreases to 10% at the possible 6ka tephra. Subsequently LOI is about 20%, and continues at this level to the core top.

Additional LOI analysis at 850°C marks the presence of increased carbonate at 770-740 cm, 700-650 cm, and 380 to 340 cm. The core stratigraphy indicates the 770-740 cm and 700-650 cm peaks are due to bivalve shells, but the

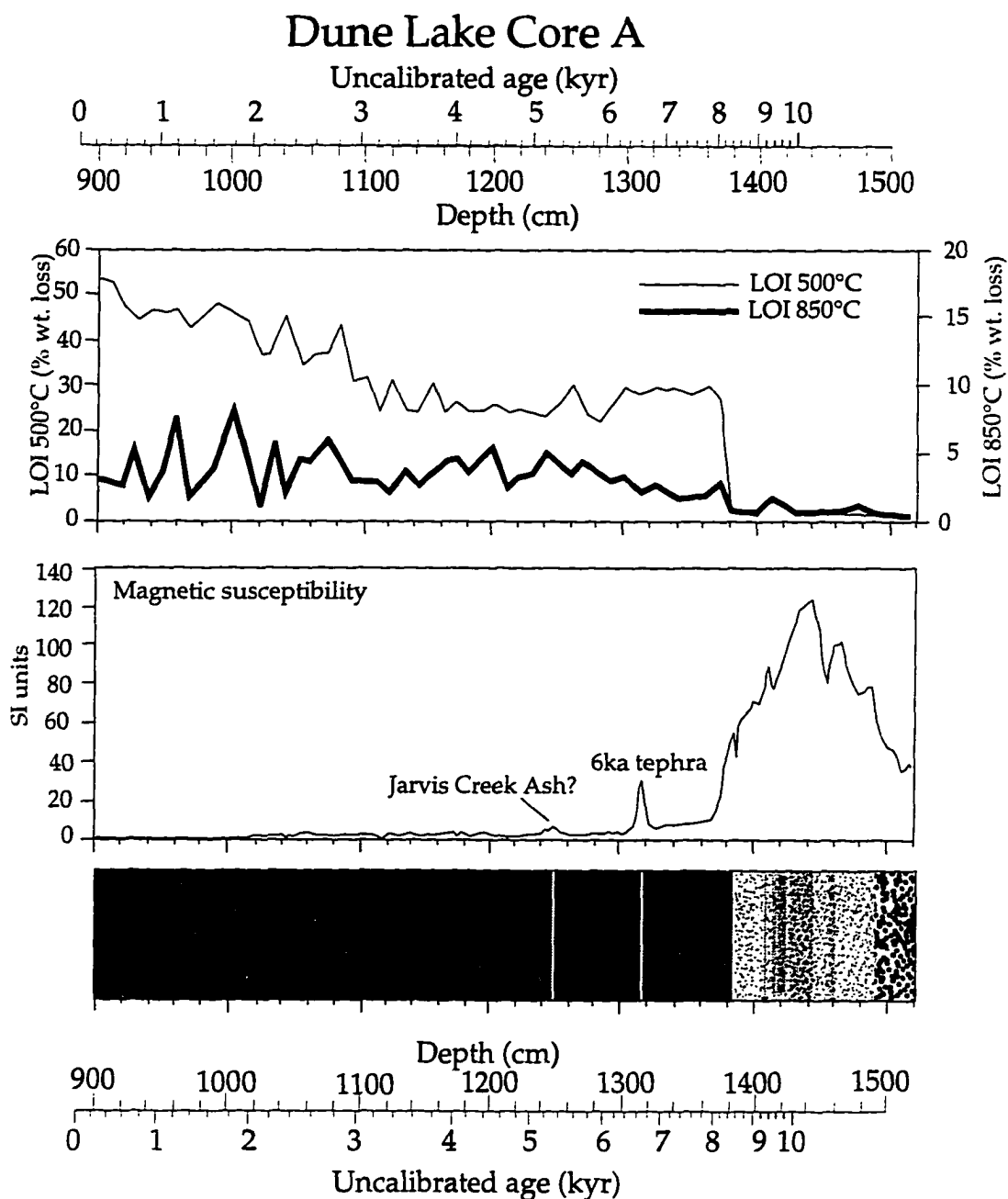


Figure 3.7: Dune A stratigraphy, LOI, and magnetic susceptibility. See Figure 3.3 for key to stratigraphic patterns.

Dune Lake Core C

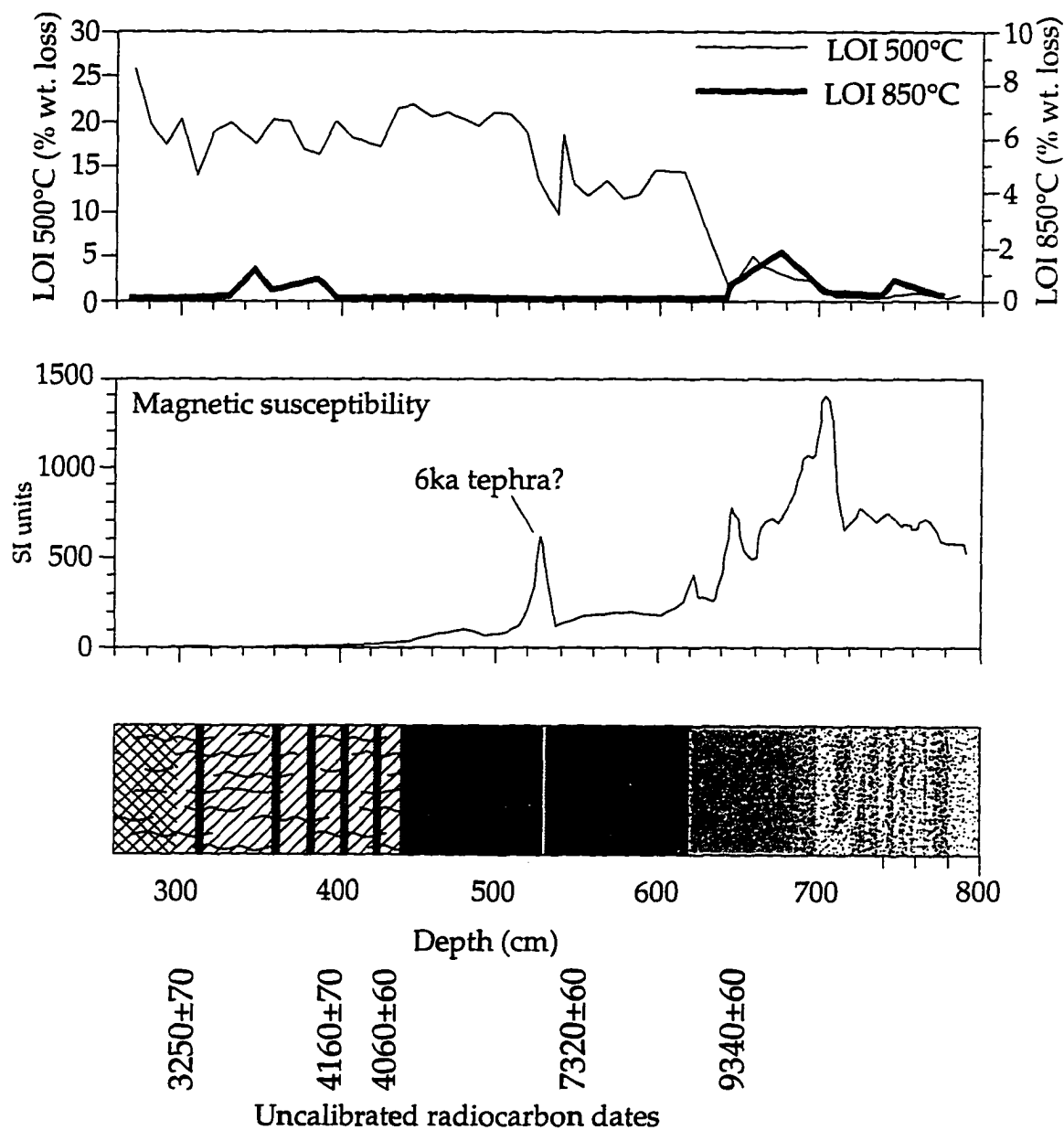


Figure 3.8: Dune C stratigraphy, LOI, and magnetic susceptibility.
See figure 3.5 for key to stratigraphic patterns.

younger peak is probably due to bicarbonate precipitation from photosynthesis.

MS analysis documents the presence of highly inorganic sediments in the base of Dune C, as well as the probable occurrence of the 6ka tephra.

Macrofossils. Preliminary macrofossil analysis was conducted on Dune C (Figure 3.9) by the participants of Dr. Hilary Birks' 1995 short-course on macrofossil identification held at the University of Alaska Fairbanks. Only the seeds have actual counts, the other remains were qualitatively measured as abundant, frequent, occasional, or rare.

From the core base to about 500 cm, macrofossils are uncommon, being limited to two levels of ostracods and mossy fragments between 700 and 650 cm. Between 500 and 450 cm, macrofossil concentration is still low, although 2 *Betula* seeds and 1 *Carex* seed were recovered.

Between 440 and 320 cm, the macrofossils are dominated by aquatic macrophyte remains. *Ceratophyllum* fragments become abundant about 4600 ^{14}C yr BP (440 cm), while high quantities of *Myriophyllum* and *Potamogeton* fragments are preserved at 355 cm and 320 cm, respectively. In addition, oospores from the hard-water loving alga *Chara* are also abundant at 320 cm.

The top sample (290 cm), marks another change, where macrofossils from the preceding submerged taxa all but disappear, and taxa from shoreline habitats (*Ranunculus scleratus*, *Rorippa islandica*, and *Carex* spp.) dominate the macrofossil assemblage.

Pollen zones. Pollen analysis was conducted on both Dune A and Dune C, but the sample interval is smallest on Dune A, thus the discussion focuses on this core. In addition, because the Dune A core lacks the top 15 cm of sediment, two samples from one of the surface cores were added to the top of Dune A to provide pollen data for the past 1000 years.

Comparison of the computer-generated zones of the actual and randomized datasets from Dune A indicate nine (CONISS) or seven (SPLITLSQ and SPLITINF) significant zones (Figure 3.10). All the computer-generated zones identified boundaries at about 8200 ^{14}C yr BP (when *Picea* appears), about 5300 ^{14}C yr BP (changes in the relative abundances of *Picea*, *Betula*, and *Alnus*-), and about 2700 and 1700 ^{14}C yr BP (fluctuating *Picea* frequencies). The pollen zones used for this analysis followed the computer-

Dune Lake Core C macrofossils

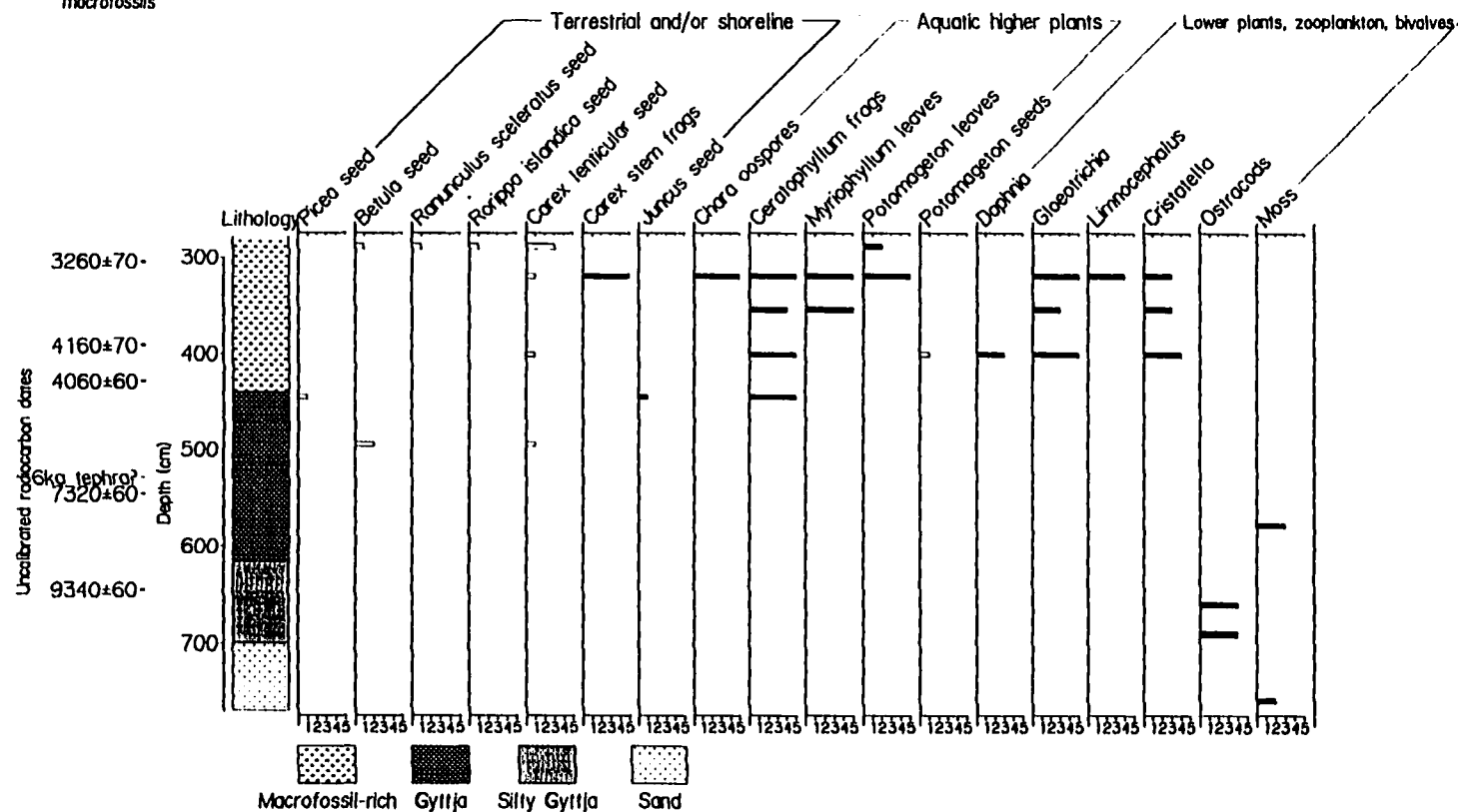


Figure 3.9: Dune C macrofossils. Open bars indicate numbers of seeds. Filled bars indicate qualitative amounts. 0 = absent, 2 = rare, 3 = common, 4 = frequent, 5 = abundant.

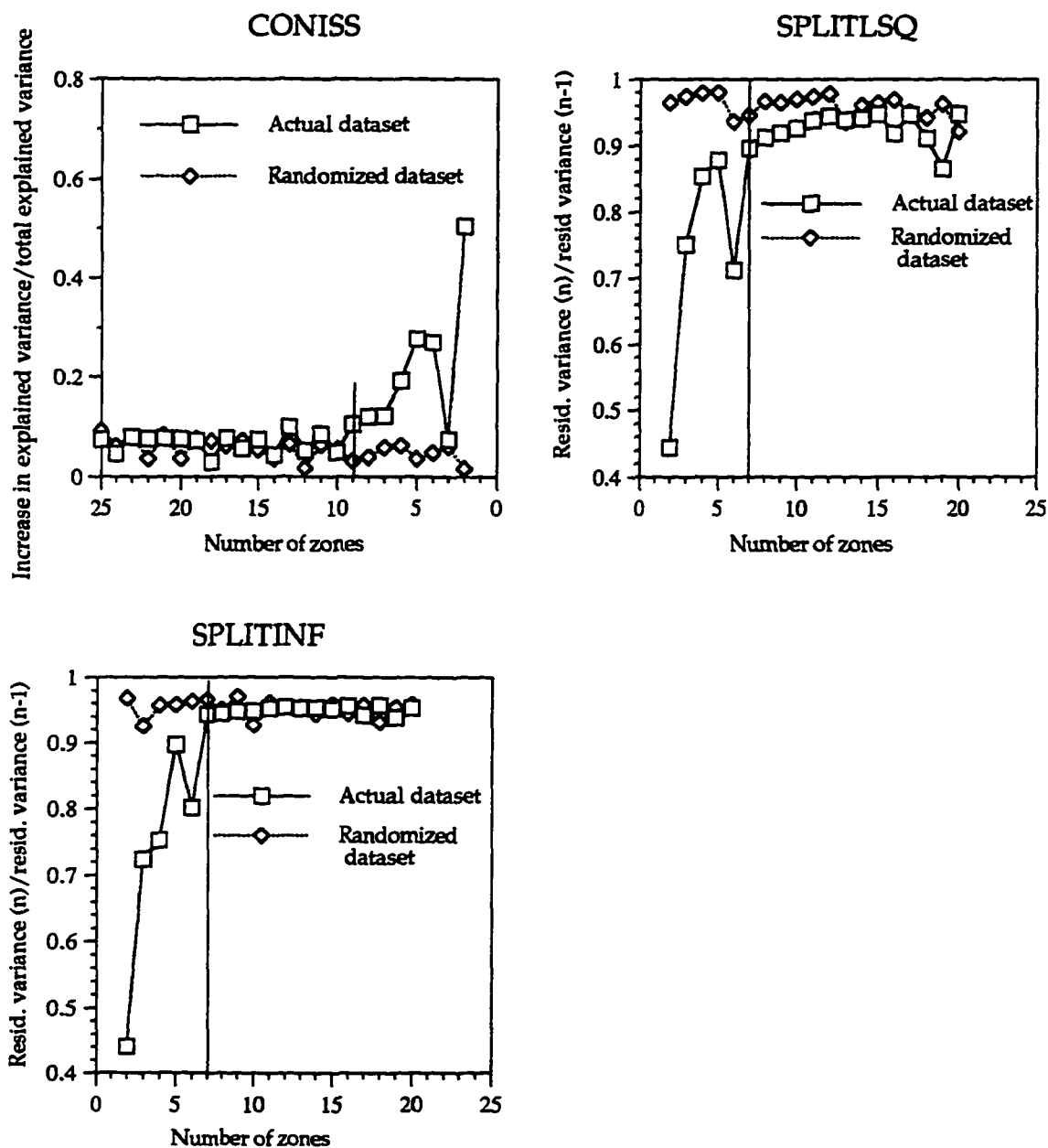


Figure 3.10: Dune core A, zoning of actual and randomized datasets, lines mark where datasets converge or diverge.

generated zones (Figure 3.11), except for the boundaries at 2700 and 1700 ^{14}C yr BP which mark subzones, because they reflect less significant changes in the vegetation than elsewhere in the core.

Pollen zone DL-1 extends from >10,000 ^{14}C yr BP to 8400 ^{14}C yr BP (1458 cm [the lowest pollen-bearing sample] to 1378 cm) (Figure 3.12). The sediments in this zone alternate between pure sand and mixtures of sandy silt and silty sand. The finer-grained sediments contain more organics and have a significantly higher pollen concentration than the sandy sediments. Pollen was virtually absent in the sediments below 1458 cm, although the presence of *Betula* indicates that the herb zone is not represented in this core. Pollen zone DL-1 is characterized by high percentages of *Betula*, *Salix*, *Populus*, *Cyperaceae*, and *Poaceae* (Figure 3.12). Total influx, however, is very low, indicating the landscape was only sparsely vegetated. The low *Betula* influx, as well as the moderate *Salix*, *Populus*, and herbaceous influx indicate that where vegetated, the landscape was dominated by a shrubby *Salix* and scattered *Populus* threes, with an understory of grass, sedge, and *Artemisia* (Fig. 3.13).

Pollen zone DL-2 extends from about 8400 ^{14}C yr BP to 7200 ^{14}C yr BP (1378 cm to 1335 cm) (Figures 3.12 and 3.13). This zone is characterized by moderate *Picea* (mainly *P. glauca*), high *Betula*, with low *Salix*, *Populus*, *Cyperaceae*, and *Poaceae* percentages. Influx analysis indicates high *Betula* influx, decreasing *Salix* and *Populus* influx, and increasing influx of herbaceous taxa. The higher quantities of *Betula* indicate this genus covered much of the landscape, while *P. glauca* was present only in low numbers. *Salix* and *Populus* were rapidly declining to modern values, which suggests that they probably only grew on the most favorable sites. The increasing influx of the herbaceous taxa (*Cyperaceae*, *Poaceae*, *Artemisia*, *Rosaceae*) indicate these taxa covered more of the landscape than during the previous zone, and that they were probably as common as they are today.

Pollen zone DL-3 extends from approximately 7200 to 5300 ^{14}C yr BP (1335 cm to 1250 cm) (Figures 3.12 and 3.13). It is characterized by increasing *Picea* (mainly *P. mariana*), decreasing *Betula* and the abrupt appearance and stabilization of *Alnus* percentages. Percentages of herbaceous taxa remain low throughout the zone. Influx analysis also indicates increasing *Picea*, but

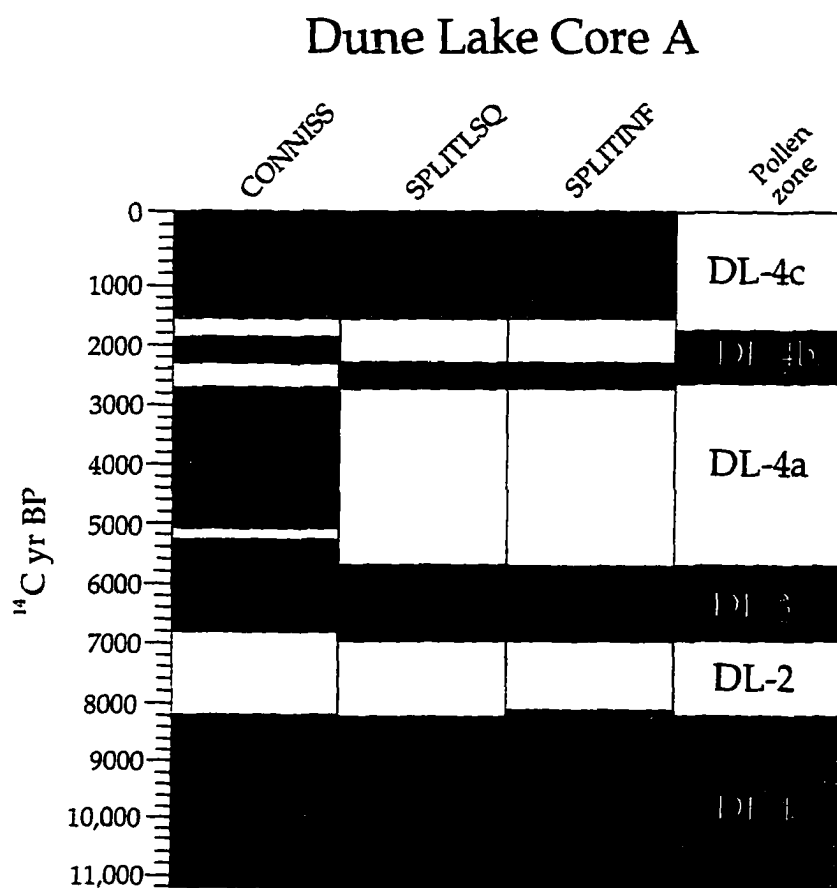


Figure 3.11: Dune Lake core A pollen zones.

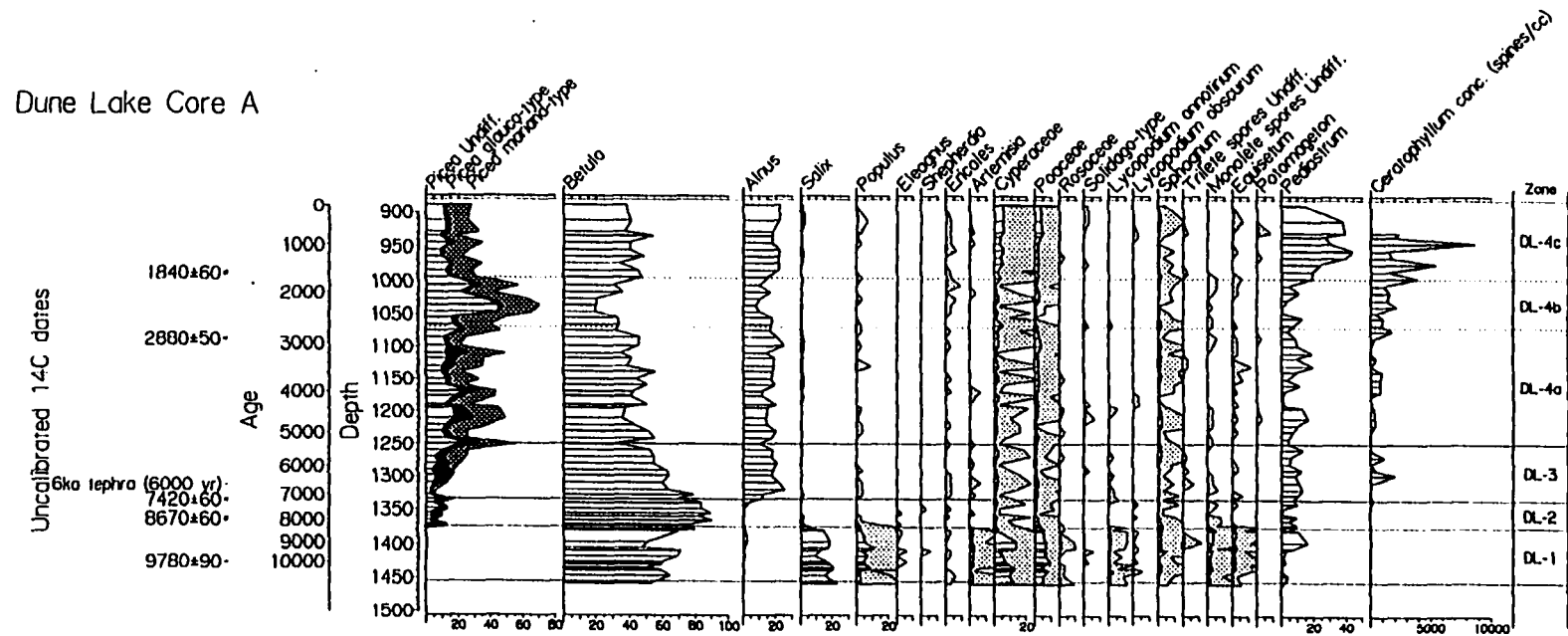


Figure 3.12: Dune Lake A pollen percentage (shading = 10x exaggeration).

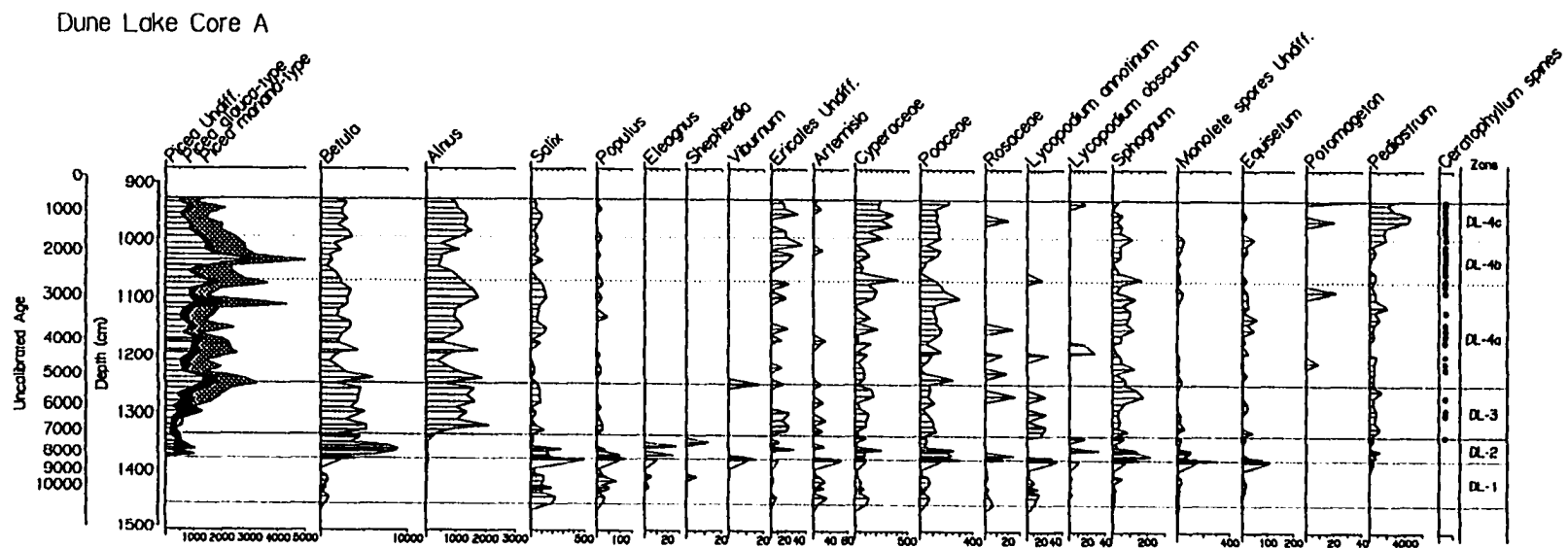


Figure 3.13: Dune Lake A pollen influx (grain/cm²/yr).

Betula influx abruptly decreases, while herbaceous influx is slightly lower than for zone DL-2. The vegetation during this zone continued to be dominated by *Betula*, but *Picea* (especially *P. mariana*) and *Alnus* were also important constituents. *Salix* and *Populus* were present only in scattered areas. Increases in the *P. mariana* percentage and influx probably indicate the initial development of the boreal forest, although it was not fully developed until the succeeding pollen zone.

Ceratophyllum cf. *demersum*, a submerged rootless aquatic, makes its initial appearance at the beginning of zone DL-3. Spines from the leaves apparently preserve well in lake sediments and are easily counted in pollen preparations. The *Ceratophyllum* is presented here as concentration data (spines/cm³) because the taphonomy of these spines is probably quite different from the pollen grains.

Pollen zone DL-4, consisting of three subzones, extends from about 5300 ¹⁴C yr BP to the present (Figures 3.12 and 3.13). Subzone DL-4a (approximately 5300 to 2800 ¹⁴C yr BP, 1250-1070 cm) is somewhat variable, with fluctuating *Picea*, *Betula*, and *Alnus* frequencies and influx, but without any overall trend. Subzone DL-4b (approximately 2800 to 1800 ¹⁴C yr BP, 1070-995 cm) is marked by high *Picea* frequencies and influx, along with lower *Betula*, *Alnus*, and *Salix* influx, suggesting increased *Picea* presence on the landscape. Subzone DL-4c (approximately 1800 ¹⁴C yr BP to the present, 995 cm to the core top) is somewhat similar to subzone DL-4a with variable *Picea*, *Betula*, and *Alnus* frequencies, but influx analysis indicates that *Picea* and *Alnus* influxes decline throughout the subzone, suggesting these species have become less common during the past 1800 ¹⁴C years. *C. demersum* concentrations increase throughout much of pollen zone DL-4, indicating this hard-water taxon (Fassett, 1957) grew well in Dune Lake during the mid- to late Holocene.

Rate of change. Rate of change of the pollen data was calculated using chord distance as the dissimilarity coefficient (DC), on the taxa listed below (Table 3.5). Ericales and Rosaceae, while not fitting the parameters for taxa inclusion outlined in Appendix A, were here used because 1) Ericales occurs frequently (although in low abundances) and 2) Rosaceae help define zone DL-2.

Table 3.5: List of taxa included in rate of change analysis.

<i>Picea</i> , undifferentiated
<i>P. glauca</i>
<i>P. mariana</i>
<i>Betula</i>
<i>Alnus</i>
<i>Salix</i>
<i>Populus</i>
Ericales
<i>Artemisia</i>
Cyperaceae
Poaceae
Rosaceae

The rate of change analysis indicates high rates of change near the core base (Figure 3.14). The peaks at 10,300 and 9200 ^{14}C yr BP reflect changes in *Betula*, while a peak at 8300 ^{14}C yr BP reflects the appearance of *Picea*. Between about 5300 and 2400 ^{14}C yr BP, the rate of change fluctuates significantly because of variable *Picea* percentages. Subsequently, the rate of change decreases.

When the rate of change is calculated using calibrated ages, there is no significant change, except the earliest peaks occur about 1000 yr earlier.

In sum, the rate of change analysis indicates two main periods of significant change in the pollen record. They occur about 10,300 to 8300 ^{14}C yr BP, and 5300 to 2400 ^{14}C yr BP.

Ordination. Ordination of the Dune A pollen data was calculated on square root-transformed percentage data using the same taxa as the rate of change analysis, except Ericales was omitted (because of poor fit with the PCA reconstruction), and the *Picea* species were summed to a single undifferentiated *Picea* genus (because the analysis tended to give high *P. glauca* loadings on the second axis; this was hard to justify, given the uncertainty in identifying the pollen grain). The gradient length of axis 1 is 1.745 standard deviation units, therefore Principle Components Analysis (PCA) was used as the ordination technique (Ter Braak and Prentice, 1988).

Dune Lake Core A

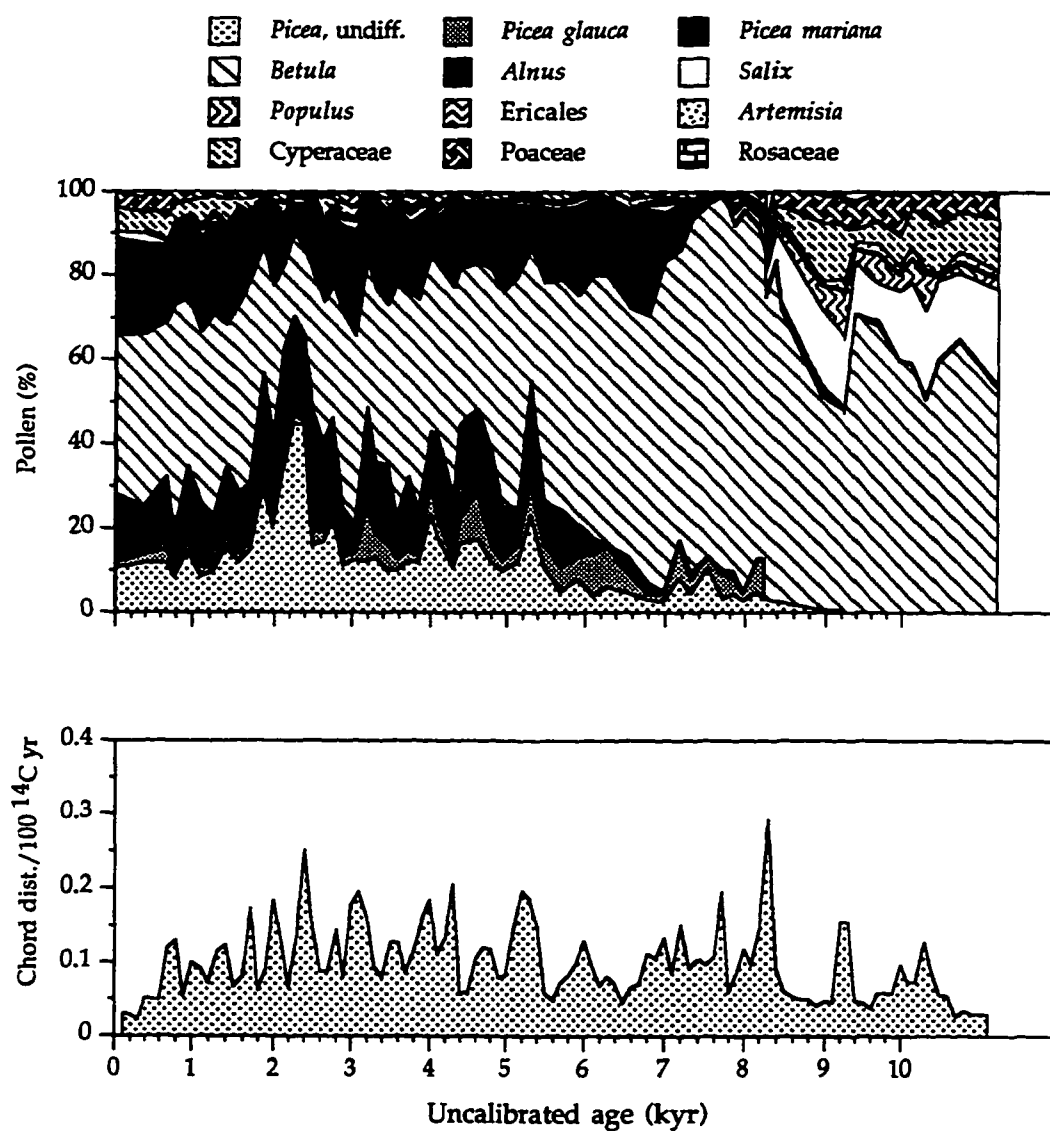


Figure 3.14: Dune A rate of change; uncalibrated age scale.

PCA axes 1 and 2 explain 75.4% and 12.1%, respectively, of the variation. The PCA biplot of the sample scores identifies the main divisions in the pollen data, although pollen zones DL-3 and DL-4 are clustered tightly together (Figure 3.15). Pollen zone DL-1 has high loadings on axis 1 and moderate loadings on axis 2. Zones DL-3 and DL-4 have low loadings on axis 1 and moderate loadings on axis 2. Zone DL-3 is intermediate between zones DL-2 and DL-4. The species scores suggest zone DL-1 is influenced by *Betula*, *Populus*, and the herbaceous taxa. Zone DL-2 is strongly controlled by *Betula*, and zones DL-3 and DL-4 are influenced mainly by *Picea*, and *Alnus*.

The loadings of the individual species on the first two axis indicate *Picea* and *Alnus* have strong negative loadings on axis 1, while *Betula*, *Salix*, *Populus*, and the herbaceous taxa have strong positive loadings (Table 3.6 species loadings). On axis 2, *Salix*, *Populus*, and the herbaceous taxa have strong negative loadings, while *Betula* has strong positive loadings.

Table 3.6: PCA species loadings.

Species	Axis 1	Axis 2
<i>Picea</i> undiff.	-113	-29
<i>Betula</i>	77	209
<i>Alnus</i>	-95	-8
<i>Salix</i>	99	-130
<i>Populus</i>	94	-108
<i>Artemisia</i>	88	-78
Cyperaceae	68	-182
Poaceae	83	-113
Rosaceae	67	-90

C and N concentrations and isotopes. Measurements of C and N concentrations and isotopes from Dune A were conducted by Bruce Finney, the data presented below are from his research. The sample interval throughout much of the core is 3 to 5 cm; to reduce some of the high-frequency variability in the dataset, the data were smoothed using a three-point running average.

C and N concentrations were measured on total sedimentary organics. At the core base, C and N concentrations are very low (Figure 3.16); these values

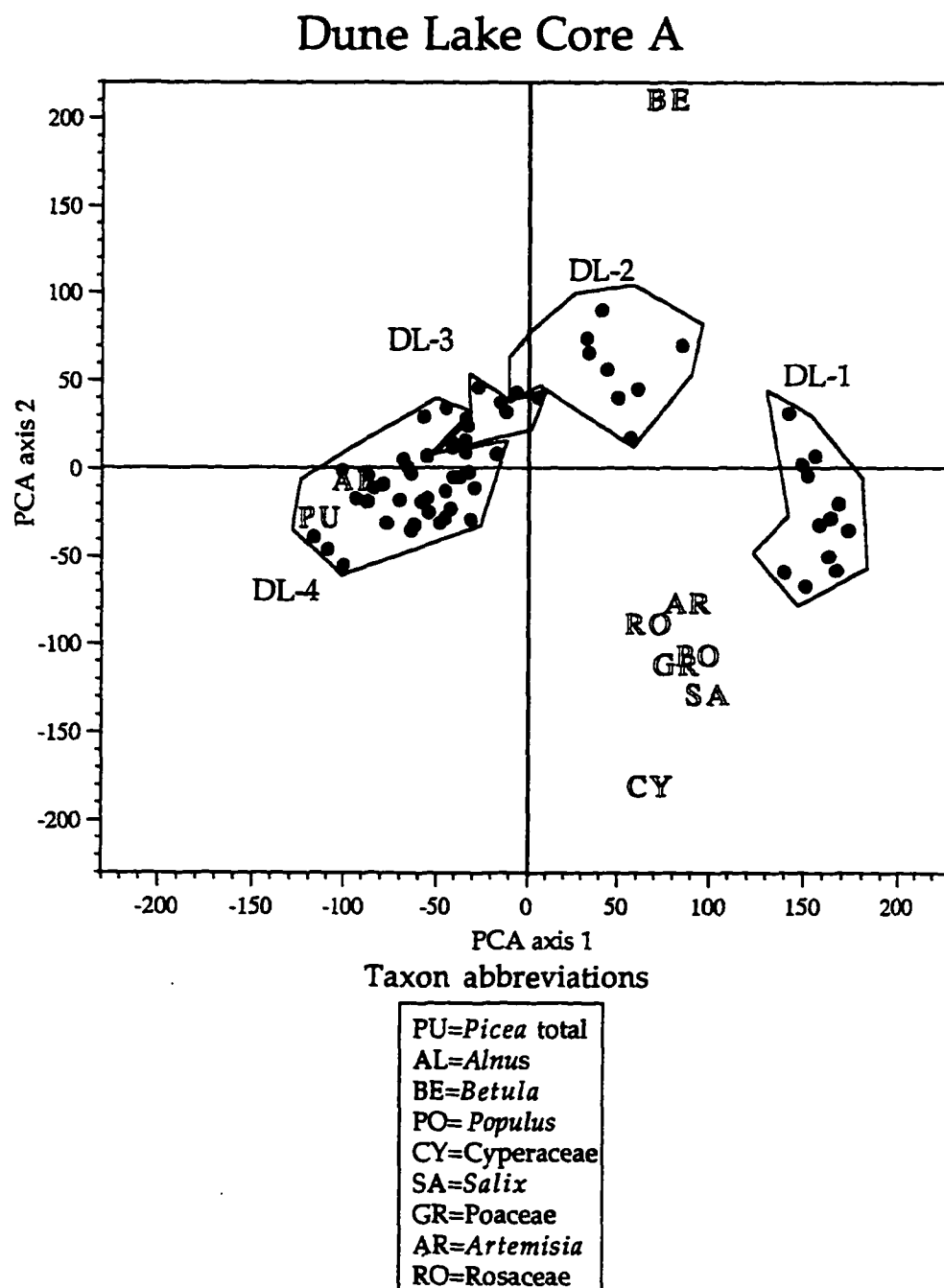


Figure 3.15: Dune A PCA biplot.

Dune Lake Core A

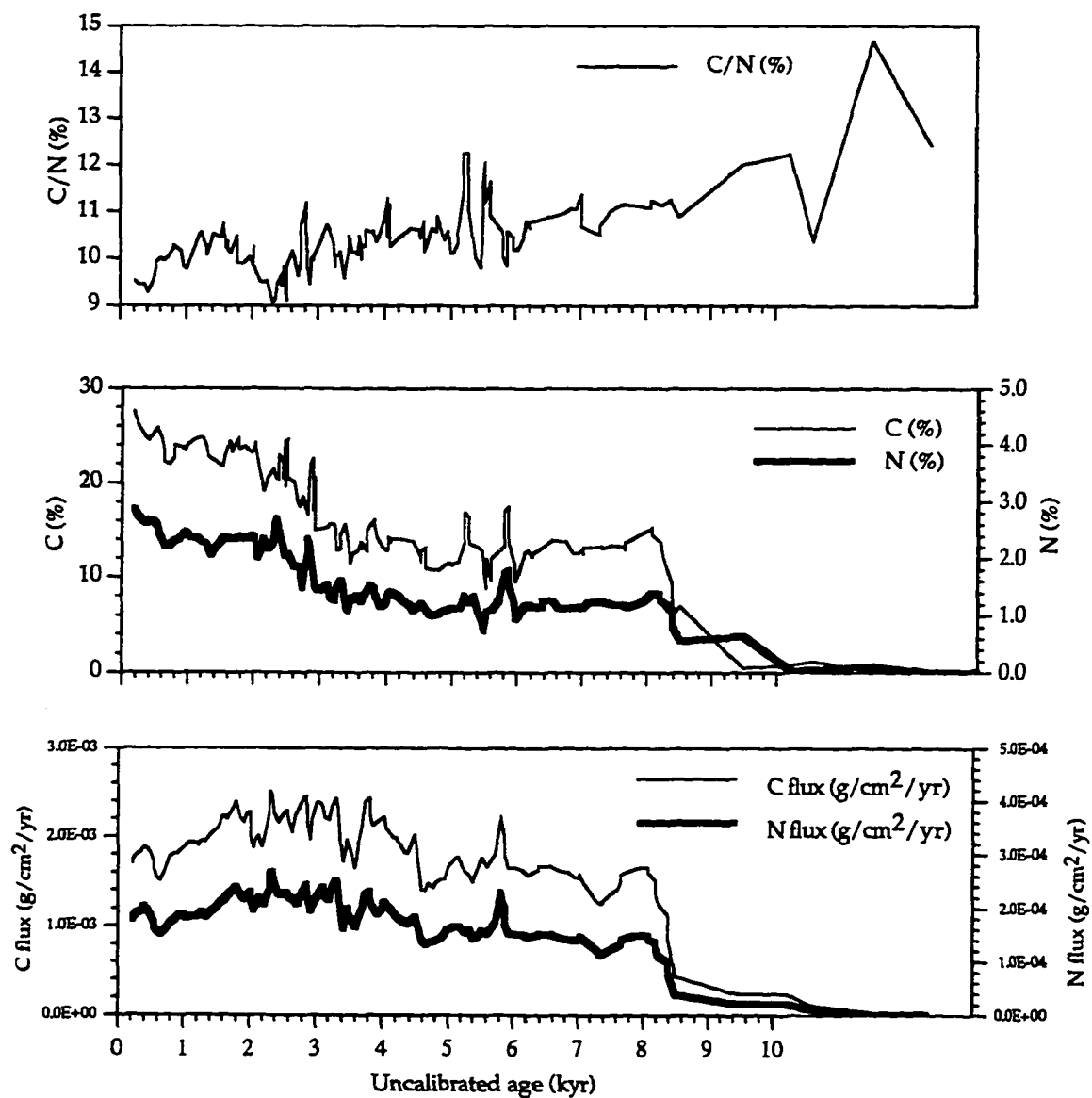


Figure 3.16: Dune A, carbon and nitrogen concentrations, C/N ratios, and fluxes.

increase dramatically at the transition to gyttja (8400 ^{14}C yr BP). Subsequently, C and N remain somewhat constant until about 3000 ^{14}C yr BP. Both C and N gradually increase at about 2800 ^{14}C yr BP, achieving modern levels by 1800 ^{14}C yr BP.

Carbon and Nitrogen fluxes ($\text{g}/\text{cm}^2/\text{yr}$) were calculated from the C and N percentage data (Figure 3.16). Both C and N fluxes are low at the core base, but increase markedly at about 8200 ^{14}C yr BP. Throughout the core, both C and N fluxes vary together, with peaks at about 5700 ^{14}C yr BP, and 5100 ^{14}C yr BP, and elevated but variable fluxes at 3300 to 1800 ^{14}C yr BP. After 1800 ^{14}C yr BP, the fluxes decrease, with a marked dip at 600 ^{14}C yr BP, and a subsequent increase to modern levels.

Most of the C and N flux reflects changes in the C and N concentrations, but the long term trends (such as gradually increasing fluxes from the core base to 2700 ^{14}C yr BP, and the subsequently decreasing fluxes to the core top) are due to the calculated sedimentation rate, which is at maximum values about 2700 ^{14}C yr BP.

The C/N ratio is high (24) at the core base, but decreases to 10-11 by about 9000 ^{14}C yr BP (Figure 3.16). Subsequently, the ratio gradually decreases until about 2000 ^{14}C yr BP, with a minimum of about 9. Between about 2000 and 200 ^{14}C yr BP, the C/N ratio is slightly elevated (10-11), subsequently decreasing to modern levels.

From the base of Dune A to approximately 10,700 ^{14}C yr BP, $\delta^{13}\text{C}$ remains fairly constant at about -26‰ (Figure 3.17). Subsequently, the $\delta^{13}\text{C}$ decreases to about -31‰ at 9600 ^{14}C yr BP, with a second, more gradual decrease to about -32.5‰ by 7200 ^{14}C yr BP. The ^{13}C then gradually increases throughout the remainder of the Holocene, reaching modern values about 2000 ^{14}C yr BP.

$\delta^{15}\text{N}$ also shows a transition near the core base, from very low values (-2‰) to high values (3‰) (Figure 3.17). The $\delta^{15}\text{N}$ then remains fairly constant at 2.5‰ between 8600 to 6200 ^{14}C yr BP. Between 5900 to 2800 ^{14}C yr BP, the $\delta^{15}\text{N}$ increases slightly to between 3.0 and 3.5‰. Subsequently the $\delta^{15}\text{N}$ decreases to about 3‰, remaining fairly constant until the core top.

Dune Lake Core A

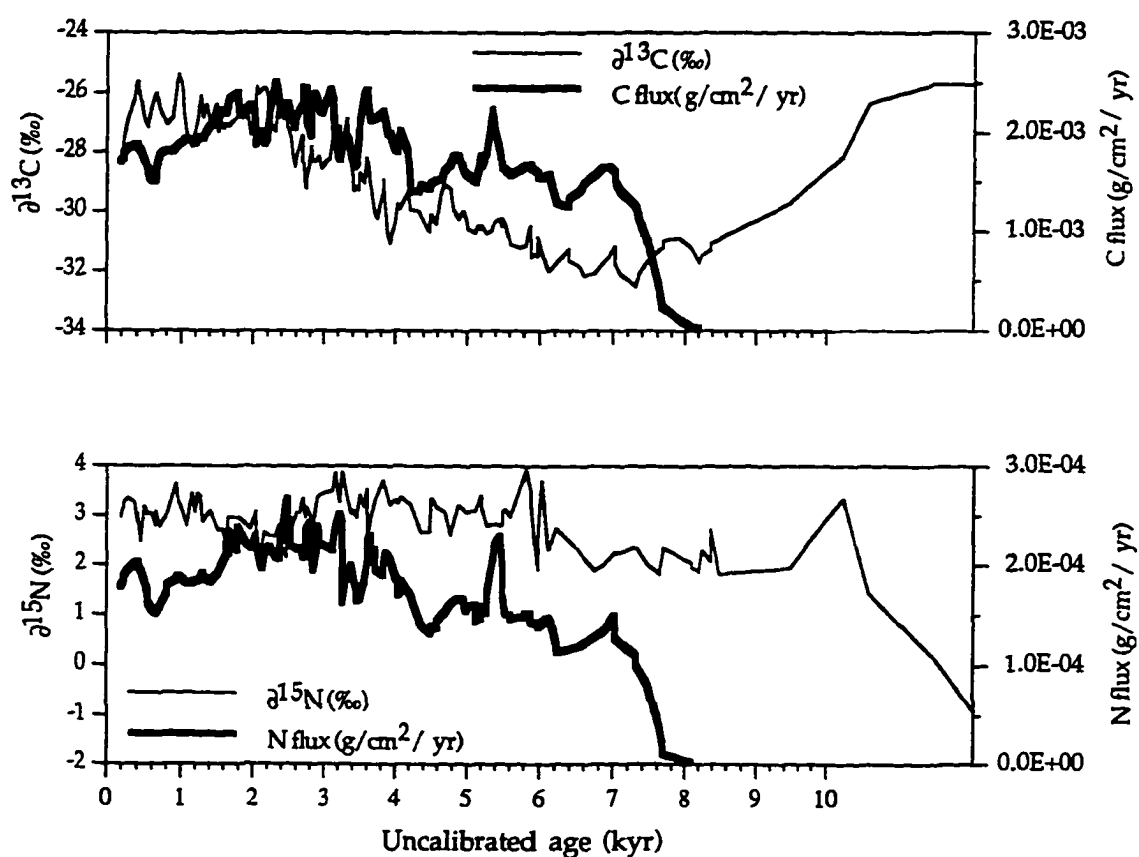


Figure 3.17: Dune core A, carbon and nitrogen isotopes and fluxes.

Discussion

Vegetation reconstructions. The pollen analysis from Dune A indicates the earliest pollen zone (>9800 to 8300 ^{14}C yr BP) is characterized by abundant *Betula*, but with significant quantities of *Salix* and *Populus*, along with moderate abundances of herbaceous taxa. Influx analysis indicates the landscape was only sparsely vegetated and was dominated by large areas of moving sand.

The sparse vegetation around Dune Lake contrasts starkly with Windmill lake, where influx analysis indicates significant coverage by *Betula* during this time period (Figure 2.9). This difference between the two lakes probably reflects local differences in soil stability and moisture availability.

The sand dunes surrounding Dune Lake are also very well drained. It is probable that the relatively high *Populus* and *Salix* abundances may be related to the warm, well-drained dune slopes. Both genera probably prospered on such sites, as well as near the lake margin. It is likely (as it is today), that *Betula* was limited to only the moister sites, particularly near the lake margin.

Usually, the *Populus* at this time period is assumed to be *P. balsamifera* because this taxon was identified (based on slight differences in the pollen grains) in a lacustrine record in the central Brooks Range (Brubaker et al., 1983). Today, however, *P. tremuloides* is common at Dune Lake, suggesting the pollen may in fact represent this species. *P. tremuloides* grows well on sunny, well drained sites (Viereck et al., 1992), the dune slopes may have been perfect habitat for this species at 10,000 to 8000 ^{14}C yr BP.

Low quantities of *Elaeagnus* and *Shepherdia* pollen are also preserved in the pollen record. These taxa are very rarely seen in pollen records, and their presence at Dune Lake suggests they may have been abundant at this time. Today, *Shepherdia* is often found on gently sloping south-facing slopes that are dominated by *Populus tremuloides* (Viereck et al., 1992). *Elaeagnus* is found on very well-drained, usually disturbed sites, such as active flood plains, road sides, or as in the case of Dune Lake, on dune slopes (Viereck et al., 1992). The presence of these taxa suggest that locally, the conditions were very dry around Dune Lake.

Pollen influx at Dune Lake increased markedly after 8400 ^{14}C yr BP. This change also apparently coincides with a major sedimentological shift from

sandy and silty sediments to lacustrine gyttja. It is unlikely that the higher pollen influx is due only to the onset of lacustrine sedimentation (and hence, improved pollen preservation), because the main influx increase in the *Picea* and *Betula* occurs about 10 cm above the transition to lacustrine gyttja (Figure 3.4).

From about 8300 to 7200 ^{14}C yr BP, pollen analysis at Dune Lake indicates the landscape was covered by abundant *Picea* (mainly *P. glauca*) and *Betula*. While the *Picea* identifications include numerous undifferentiated pollen grains, the dominance of *P. glauca* during the early Holocene has been repeatedly demonstrated at numerous other lakes (Anderson and Brubaker, 1994).

In central Alaska, when quantitative spruce measurements have been made, the initial appearance of *Picea* is usually marked by the dominance by *P. glauca*, with *P. mariana* only becoming abundant later in the mid-Holocene (Hu et al., 1993; Hu et al., 1996; Anderson et al., 1990; Anderson et al., 1988; Edwards and Brubaker, 1986). One possible exception is Seagull Lake in the eastern Brooks Range (Figure 1.2), where *P. mariana* apparently is present by 8000 ^{14}C yr BP (Anderson and Brubaker, 1994).

In most sites in western Canada, wherever *Picea* identifications have been made, the earliest spruce has been *P. glauca*, indicating the post-glacial spruce migration was made mainly by white spruce, with black spruce trailing along behind (Ritchie and MacDonald, 1986). However, one site in the southern Yukon Territory (Long Last Lake) (Figure 1.2) indicates that *P. mariana* may have been present as long ago as 8000 ^{14}C yr BP, appearing locally only 500 ^{14}C yr after the arrival of *P. glauca* (Keenan and Cwynar, 1992).

Analyses of pollen cores from south of the Laurentide ice sheet indicate that *P. glauca* was the dominant species closest to the ice edge (Ritchie and MacDonald, 1986). Possibly the *P. glauca* migrated fastest because it occupied strategically important sites and could colonize the newly available land ahead of the *P. mariana*. The rate of migration is impossible to measure because of uncertain chronologies, that make it appear that *Picea* arrived simultaneously across large areas of Canada and eastern Alaska (Ritchie and MacDonald, 1986). In addition new AMS dates on the *Picea* rise tend to be significantly younger than the conventional dates for this event.

Simultaneously with the arrival of *P. glauca*, *Betula* influx increases to greater than modern values at Dune Lake. Wien Lake also has high *Betula* influx at this time (Hu et al., 1993), but most other lakes in central Alaska record highest *Betula* influx prior to the arrival of *Picea*. Macrofossil analysis at Wien Lake indicates the local arrival of *B. papyrifera* and the influx increase occurred simultaneously, suggesting the high influx reflects the addition of *B. papyrifera* to the local vegetation, although an alternative hypothesis is the high *Betula* influx also reflects a change to more mesic conditions, allowing *Betula* as well as *Picea* to colonize new sites.

By about 6800 ^{14}C yr BP, *Alnus* frequencies cross the 20% threshold, indicating it probably grew locally, but was not common (it is not common at Dune Lake today). Analyses of modern vegetation/pollen relationships suggest *Alnus* pollen over-represents the vegetation, and that the 20% frequency indicates probable local presence (Anderson and Brubaker, 1986).

Alnus abundances tend to be significantly higher in western and southwestern Alaska than in central Alaska. At Windmill, Dune, and Wien Lakes, *Alnus* fluctuates at around 20%, suggesting that while present, the genus did not dominate the landscape. In contrast, lakes in the western Brooks Range (Joe, Niliq, and Kaiyak lakes [Anderson, 1988; 1985]) and southwest Alaska (Grandfather and Ongivinuk lakes [Hu et al., 1995]) often have 40 to 60% *Alnus* pollen, suggesting it grew very commonly, probably reflecting more mesic conditions there.

The earliest postglacial *Alnus* pollen is preserved in sites in the western Brooks range (Anderson and Brubaker, 1994), suggesting this genus could have survived the full glacial in refugia in western Alaska or on the Bering Land Bridge. Because of poorly constrained chronologies, it is difficult to trace the *Alnus* migration, although, as a rule, *Alnus* tends to appear later at sites in the east from about 6000 to 68000 ^{14}C yr BP, while in the northwest, it appeared about 8000 to 9000 ^{14}C yr BP (Anderson and Brubaker, 1994).

From about 5500 to 2000 ^{14}C yr BP, *Picea* percentage and influx values fluctuate markedly at Dune Lake, with the *Picea*, *Betula*, and *Alnus* influxes increasing and decreasing together. Farewell lake may show some evidence of the same pattern (Hu et al., 1996). The nearby Wein Lake does not, but this may be because the sample interval (ca. 500 ^{14}C yr) is too large. The Dune Lake

pattern is in contrast to Windmill lake, where *Picea* increases when *Betula* and *Alnus* decrease. This suggests that Dune Lake (and possibly Farewell Lake) are recording a local phenomenon which was not reflected at other sites in central Alaska.

At Dune Lake, the three taxa increase about 5300, 4400, 3700, and 3200 ^{14}C yr BP, while decreasing about 4800, 4000, and 3400 ^{14}C yr BP. The time interval between low and high influxes is 200 to 500 ^{14}C yr (Figure 3.18). This interval is a little long for forest regeneration after a forest fire (Dyrness et al., 1986), and charcoal was not common in the pollen core. With the exception of *P. glauca*, all the species are mesic taxa, indicating these fluctuations may reflect short-term increases in effective moisture, although *Salix* (a mesic taxon) varies independently. Herbaceous taxa do not show any obvious patterns, although they tend to have higher influxes when *Picea*, *Betula* and *Alnus* have high influxes. These data suggest that episodically, conditions became more favorable for most taxa. Increasing effective moisture is the most likely candidate, because vegetation in much of the area around Dune Lake is currently moisture limited. Even a small change in effective moisture may have lead to significant changes in the local vegetation, with the *Picea*, *Betula*, and *Alnus* colonizing the currently dry south-facing dune slopes.

At about 2200 ^{14}C yr BP, *Picea* percentages and influx at Dune Lake are two to four times their modern values. During this peak, however, *Betula* and *Alnus* decrease, suggesting a different mechanism forced this change.

Higher than modern *Picea* abundances have been documented at Windmill Lake, where beginning about 5500 ^{14}C yr BP, *Picea* peaks episodically exceed modern levels. At Birch lake, *Picea* influx also episodically exceeds modern levels before 2500 ^{14}C yr BP.

These data suggest that *Picea* was more abundant around Windmill, Dune, and Birch lakes during the mid-Holocene than it is today. This pattern is not commonly repeated in other central Alaskan lakes. Exceptions are Wonder Lake where mid- to late Holocene *Picea* frequencies are higher than modern (Anderson et al., 1994), and Wien Lake which shows higher than modern influx about 6000 ^{14}C yr BP, gradually decreasing to modern levels (Hu et al., 1993). However, most lakes show that if *Picea* frequency or influx ever approaches modern levels, this most often occurs in the early Holocene,

Dune Lake Core A

influx of selected species

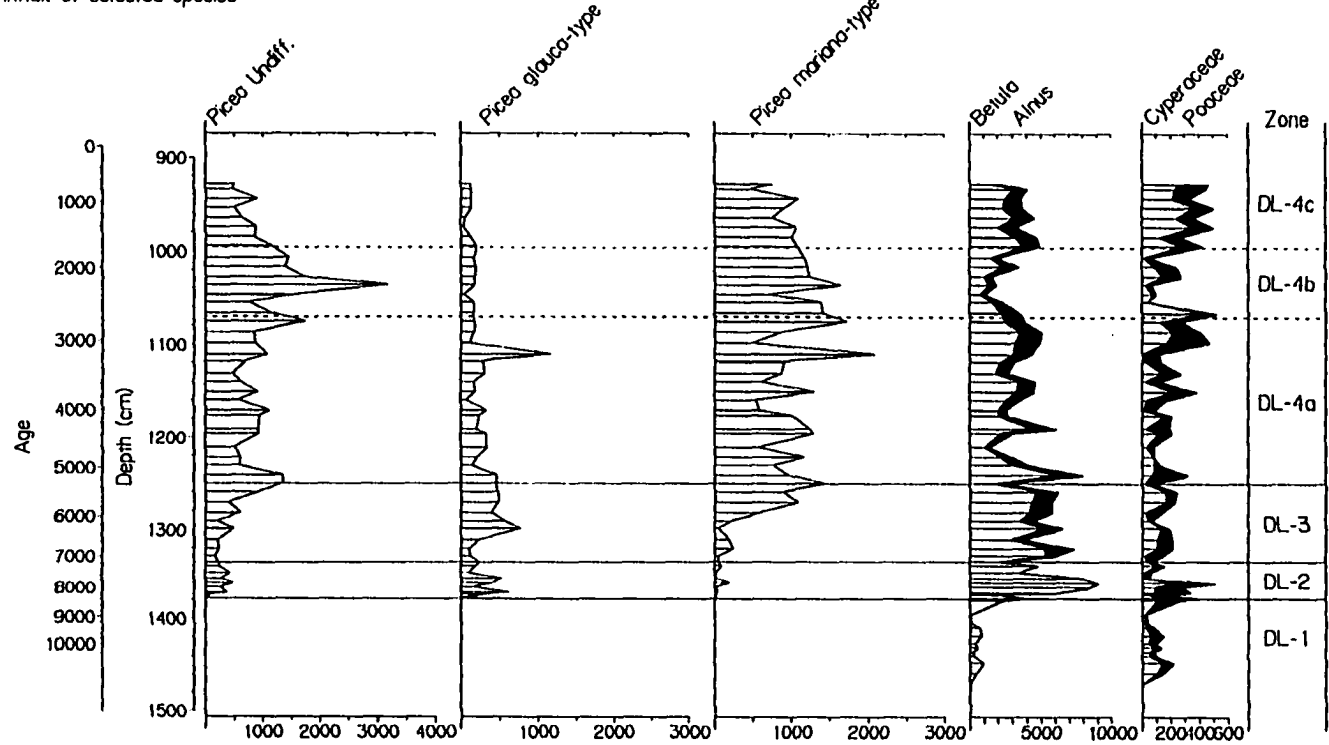


Figure 3.18: Dune A influx of dominant taxa.

during the initial *Picea* migration into the region (Anderson et al., 1990; Anderson et al., 1994; Edwards et al., 1985).

Pollen influx data from Dune Lake imply that *Picea* (mainly undiff. and *P. mariana*) expanded and occupied areas where *Betula* and *Alnus* formerly grew. This may have been due to the continuation of higher effective moisture, which was beginning to favor the most cold-tolerant and mesophytic species over those less tolerant of cold wet soils.

Alternatively, this increase in *P. mariana* may also reflect successional changes where areas of *Betula* and *Alnus* woodland developed into a *P. mariana* forest and/or bog because the *Betula* and *Alnus* were less able to withstand cooler soil temperatures resulting from incremental additions of organic matter on the ground surface (Viereck, 1970).

It is also possible that the high *Picea* influx reflects the onset of cooler atmospheric temperatures, that initially favored *Picea* over *Betula* and *Alnus*. Whatever the cause for the high *Picea* influx 2000 to 4000 ^{14}C yr BP, *Picea* declined rapidly in central Alaska after 2000 ^{14}C yr BP.

It is not clear whether the *Picea* influx decline reflects changing *P. glauca* or *P. mariana* distributions. At Dune Lake, most of the change is in the *Picea* undiff. group. *P. glauca* is constant at low levels, while *P. mariana* declines. However, because of the difficulties in reliably identifying *P. glauca* and *P. mariana* pollen, not too much weight should be put on this. At Wien Lake, quantitative analysis of the *Picea* grains indicates most of the decline is in the *P. glauca*, although *P. mariana* decreases as well (Hu et al., 1993). At Dune Lake, while *Picea* is declining, most of the remaining taxa are unchanged, although *Alnus* decreases slightly, while *Cyperaceae* increases. At Wein Lake, all taxa decrease, although *Sphagnum* increases slightly at the end of the record (Hu, et al., 1993).

It is possible that the declining *Picea* reflects only successional changes (Viereck, 1970), but continue climatic deterioration during the Late Holocene was probably an important factor. Late Holocene climatic deterioration has been suggested by renewed ice sheet expansion (Hamilton, 1986b; Calkin, 1988), lowered tree-lines in northwestern Canada (Ritchie et al., 1983; Moser and MacDonald, 1990), and ice wedge growth (Hamilton et al., 1983). Cooler summers would have resulted in increased effective moisture, reduced active

layer thicknesses and expansion of permafrost, especially in areas already dominated by *Picea*. At the onset of cooling, *P. mariana* may be favored over *P. glauca*, but as cooling continued *P. mariana* would have been severely reduced or eliminated from areas with poor drainage, while *P. glauca* could continue on the well-drained sites, such as hillsides, or in the case of Dune Lake, on the sand dunes.

Geochemical analyses at Farewell Lake (Hu et al., 1993), suggest that the onset of paludification may have occurred as early as 4000 ^{14}C yr BP. At Dune Lake, however, the *Picea* didn't decrease until after 2200 ^{14}C yr BP. This difference is probably due to the sandy soils surrounding Dune Lake, which would have limited paludification to interdunal sites and some north-facing slopes. In contrast, the landscape surrounding Farewell Lake was probably significantly more mesic than Dune Lake. Thus, with the onset of climatic cooling and increased effective moisture, widespread paludification would have started earlier there than at Dune Lake.

Stable carbon isotopes. Much of the particulate organic matter in Dune A is probably derived from within the lake. Throughout most of the core, C/N ratios are <10 , and $\delta^{15}\text{N}$ is $>2\text{‰}$, indicating relatively low fluxes of terrestrial organic matter into the lake. The exception, however, is at the core base, where C/N ratios are high and $\delta^{15}\text{N}$ is low, both suggesting reworked terrestrial material. This suggests that in the upper part of the core the carbon and nitrogen isotopes reflect changes in the lake's water chemistry but that at the core base, the isotopes may reflect other factors.

Stable carbon isotopes of aquatic-derived organic matter reflect the $\delta^{13}\text{C}$ of the DIC entering the lake; whether it originated from the atmosphere, groundwater, or from the bedrock (Peterson and Frye, 1987). In addition, the $\delta^{13}\text{C}$ of the sediments also reflect the change of the $\delta^{13}\text{C}$ of DIC within the lake. That is, during periods of high productivity, the $\delta^{13}\text{C}$ of organic carbon becomes enriched in ^{13}C because of reduced fractionation, bicarbonate production, or drawdown of atmospheric CO_2 (Gu et al. 1996) (see also Chapter 2 and Appendix B for a more detailed discussion).

At Dune Lake, the $\delta^{13}\text{C}$ signal of the lake water DIC mainly reflects the low $\delta^{13}\text{C}$ of the groundwater. The groundwater is depleted in ^{13}C because of the CO_2 respired from the C_3 surface vegetation and soil organic matter. For

most of the year, Dune Lake probably does not absorb atmospheric DIC, because most Arctic lakes are apparently supersaturated in CO₂ (Kling et al., 1991). Only abundant productivity, by depleting the DIC in the lake, could reverse this trend.

At the core base (>10,000 ¹⁴C yr BP), the δ¹³C of the lacustrine sediment is very high (Figure 3.17). C and N concentrations and fluxes are low at this time, suggesting productivity was also very low. The high δ¹³C thus probably reflects a change in the lake water DIC, possibly as a result of reduced groundwater flux into the lake. This is also supported by the lowered lake levels at this time, which reflects reduced effective moisture. Today, simple water balance equations from two central Alaskan lakes suggest that groundwater and surface run-off account for 1/2 to 2/3 of the water inputs into these lakes (B. Finney, pers. comm.). Thus, a reduction of groundwater flux would have a major effect on both the lake level, and on the δ¹³C signal of the lake water DIC, because the lake would be closer to isotopic equilibrium with the atmosphere.

From about 10,000 ¹⁴C yr BP to about 8400 ¹⁴C yr BP, δ¹³C of the lake sediments decreases markedly, C/N ratios decrease, while C and N concentrations and fluxes increase slightly (Figures 3.16 and 3.17). These changes suggest a decrease in terrestrial organic matter, and increased lake productivity, although not enough to limit carbon and affect the C isotope signal. Stratigraphic changes in the lacustrine cores suggest the lake level was rising at this time, which explains the reduced C/N ratios and higher fluxes. The lowered ¹³C of the lake sediments probably reflects increased flux of groundwater DIC into the lake, which is consistent with a higher lake level.

After 8400 ¹⁴C yr BP, the δ¹³C of the sediments increases gradually until about 3800 ¹⁴C yr BP. After 3800 ¹⁴C yr BP, the δ¹³C increases more rapidly, and then stabilizes at -27‰ to -26‰ from about 2000 ¹⁴C yr BP until the core top (Figure 3.17). Carbon concentrations and fluxes increase dramatically at 8400 ¹⁴C yr BP, but then continue to increase more gradually throughout much of the Holocene (Figure 3.16). C/N ratios decrease throughout the period (Figure 3.16). These changes suggest that once the sand dunes surrounding Dune Lake were stabilized at about 8400 ¹⁴C yr BP, significantly less terrestrial organic matter was being deposited in the lake and lake

productivity increased substantially. The increasing ^{13}C signal of the lake sediments probably reflects lowered fractionation against ^{13}C or increasing amounts of bicarbonate in the lake's DIC pool as productivity increased.

Stable nitrogen isotopes. Stable nitrogen isotopes are particularly sensitive to nitrogen cycling within the lake, as well as changes in the relative amount of terrestrial DIN in the lake's nitrogen pool (Peterson and Fry, 1987; Estep and Vigg, 1985; Gu et al., 1994; Gu, 1993c; France, 1995b) (See Chapter 2 and Appendix B for a more detailed discussion).

In Dune A, $\delta^{15}\text{N}$ is quite low and the C/N ratio is high in the sediments at the base of the core, consistent with more abundant terrestrial N (Figures 3.16 and 3.17). The lake level was very low at this time (see following section). It is probable that the nitrogen isotopes reflect not only increased terrestrial N transport at the coring site (because the coring site would have been close to the lake edge), but also very low lake productivity (because less N is transformed prior to sedimentation).

About 10,000 ^{14}C yr BP, $\delta^{15}\text{N}$ increases sharply, and remains at a stable level (about +2‰) until about 6000 ^{14}C yr BP. C/N ratios decrease, while C concentrations and fluxes increase during this time. This suggests that the higher ^{15}N content of the sediments reflects increased lacustrine organic matter, as well as higher productivity. This is consistent with the decreases in $\delta^{13}\text{C}$ during this time period which indicate higher groundwater flux into the lake (Figures 3.16 and 3.17).

Between about 6000 and 2800 ^{14}C yr BP, $\delta^{15}\text{N}$ increases (to +3‰), it subsequently decreases slightly (to +2.5‰) for a short period (about 2800-1800 ^{14}C yr BP), and then returns to modern levels (about +3‰) (Figure 3.17). These changes cannot be directly correlated with changes in C/N ratios, C concentrations, or C fluxes. Despite this, these relatively small changes in $\delta^{15}\text{N}$ may reflect changes in lake productivity. The interval of elevated $\delta^{15}\text{N}$ between 6000 and 2800 ^{14}C yr BP may reflect a period of intense lake productivity, when N utilization and recycling rates increased. Higher productivity is also supported by the higher apparent sedimentation rate, as well as strong, thick laminations in the core during the middle Holocene.

The decrease in $\delta^{15}\text{N}$ at 2800-1800 ^{14}C yr BP may reflect increased denitrification and increased use of terrestrial DIN by lacustrine organisms.

Alternatively it reflects an increase of blue-green algae. Blue-green algae do not fractionate when they fix nitrogen, thus they have a $\delta^{15}\text{N}$ signal close to the atmosphere (Handley and Raven, 1992). A period of more abundant blue-green algae in the lake water may have been sufficient to decrease the $\delta^{15}\text{N}$ of the sediments, as has been suggested for a tropical lake in Africa (Talbot and Johannessen, 1992).

Lake level changes. The presence of subaerial and submerged shorelines at Dune Lake indicate the water level has fluctuated markedly in the past. In addition, the water level increased about 3 m since the early 1970s, indicating this lake is sensitive to short-term changes to precipitation and evapotranspiration within the watershed (Barber and Finney, 1995).

Lake level changes (Figure 3.19) at Dune lake are marked by lithological and vegetational changes in cores A and C. The earliest parts of both cores are undated, so that no age assessments have been made on the pre-Holocene lake level changes

Both cores have coarse inorganic sand at their bases, indicating the lake was probably a waterless interdunal basin before 10,000 ^{14}C yr BP. Subsequently, the sand becomes finer and is interspersed by lenses of silt. The silt bands in Dune C probably reflect lacustrine deposition because of the presence of ostracods and *Pediastrum* (a green alga). Thus, the water must have been at least 7 m deep (but about 7 m below modern levels) in the central basin for it to reach the Dune C locality (Figure 3.2).

The bottom radiocarbon date from Dune A suggests the silt layers were deposited there from before 9700 ^{14}C yr BP to about 9000 ^{14}C yr BP. The sand separating the silt and sandy silt can be interpreted in two ways. The sand may represent renewed eolian deposition into the lake, or it may represent active erosion of the sand dunes at the lake margin and redeposition into the lake basin. These two models differ not only in their implied climates, but also in their implied time scales. The eolian model suggests a dry, windy regime and sparse vegetation cover, with sand deposition occurring over a number of seasons. In contrast, the redeposition model suggests increasing water levels eroded the shoreline, with the potential for very rapid sediment deposition by turbidites.

Dune Lake

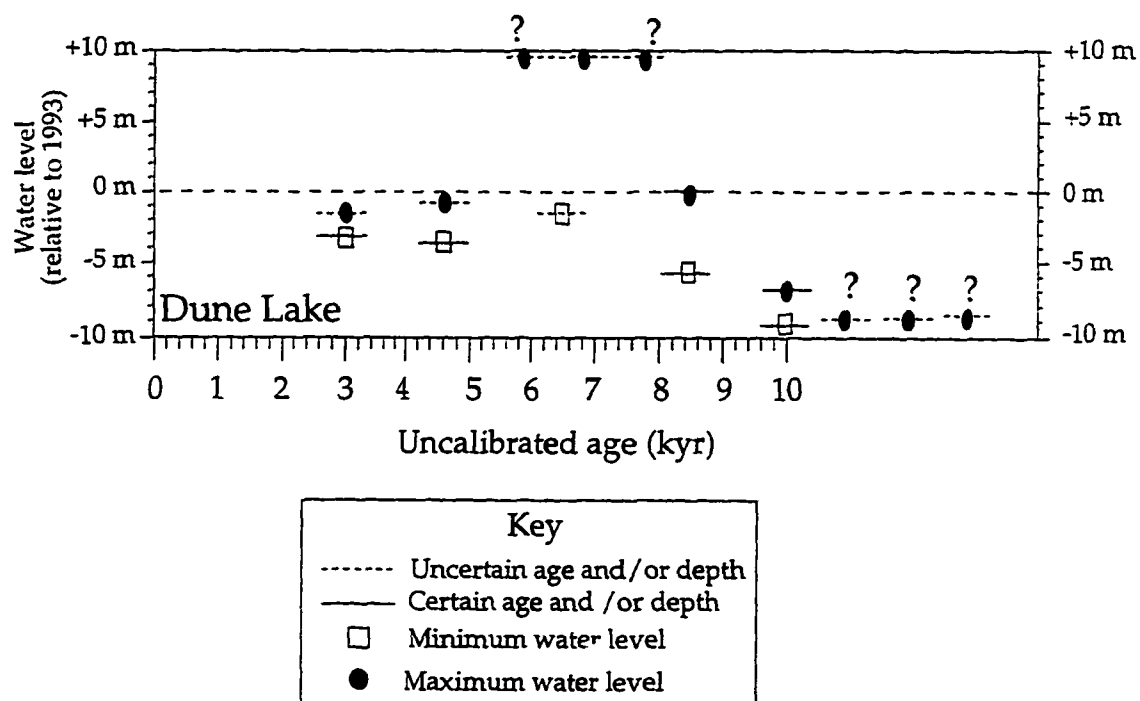


Figure 3.19: Dune Lake lake level changes.

Both mechanisms may have deposited sand into Dune Lake. The vegetation reconstruction indicates that the dunes were not stabilized and plants were quite sparse in the region, so that eolian processes could have easily deposited sand in the basin. The carbon isotopes suggest reduced groundwater flux and the nitrogen isotopes suggest abundant terrestrial material, indicating an arid landscape, with significant erosion into the basin. In addition, the complex stratigraphy in Dune C also indicates sediment reworking near the shoreline as the lake level rose.

After about 8400 ^{14}C yr BP, lacustrine gyttja dominates both cores, indicating not only a significant increase in the water level (i.e., migration of the sediment limit shoreward of the coring sites [Digerfeldt, 1986]), but also dune stabilization. Pollen influx analysis indicates the landscape was densely vegetated by about 8400 ^{14}C yr BP. This would have halted most sand movement into the lake and allowed shoreline stabilization.

Beginning about 6300 ^{14}C yr BP, strong laminae are preserved in Dune A. Laminations occur when bottom water becomes anoxic and annual additions of sediment are not bioturbated. Today, Dune Lake is anoxic during the summer below 7 to 8 m of water. The presence of the laminae suggest both that the lake was at least 7 m deep, and that productivity was high enough to produce sufficient detritus to deoxygenate the bottom water through respiration. Increasing C and N fluxes also support higher productivity at this time. The lake level was probably higher than modern, as indicated by exposed shorelines around Dune Lake. The highest shoreline indicates the lake level rose to a maximum of 10 m above modern levels.

By roughly 4000 ^{14}C yr BP, the water level was again decreasing, because in the Dune C core sediments become increasingly rich in macrofossil remains, which become coarser near the core top (indicating migration of the shoreline towards the Dune C coring site). The macrofossil remains are initially dominated by submerged taxa and zooplankton, although about 3300 ^{14}C yr BP emergent taxa become important, and by 3100 ^{14}C yr BP they are the only taxa represented.

After 3000 ^{14}C yr BP, Dune C is truncated, indicating it has been subaerially exposed since that time (although we do not know if the core top has been eroded). In Dune A, laminae are still present after 3000 ^{14}C yr BP, although

they are thinner and less continuous than in the earlier part of the core, suggesting either lower water level or reduced productivity. About 1800 ^{14}C yr BP, *Ceratophyllum* spines become more common and the C/N ratio increases slightly in Dune A, suggesting a lowered water level as the lake edge moved closer to the coring site.

Summary

The pollen, isotopic, and stratigraphic analyses indicate that somewhat before 10,000 ^{14}C yr BP, Dune Lake was a dry interdunal basin. At some point thereafter, the lake level rose, depositing silty lacustrine sediment in the central basin and along the shore, suggesting the water level was about 7 m below modern. Vegetation at the end of this period (9300 to 9800 ^{14}C yr BP), was a sparse scrubby plant cover, dominated by *Betula*, *Populus*, and *Salix*.

By about 8400 ^{14}C yr BP, organic-rich lacustrine sediments were deposited in the central and side cores, indicating the water level was at the most 7 m below modern, but was probably significantly higher. At this time, *Picea* was present on the landscape, and there was near complete vegetation coverage of the ground surface.

About 7500 ^{14}C yr BP, laminations are preserved in the central core, indicating high water levels and high productivity. The water level was probably higher than modern, the raised beaches suggest it reached a maximum level of 10 m above modern. Shortly after 7500 ^{14}C yr BP, *P. mariana* and *Alnus* become important constituents of the vegetation. The high effective moisture may have promoted the dispersal of these mesic species. Perhaps by about 5000 ^{14}C yr BP, lake levels began to decrease again, so that Dune C was near the shoreline or subaerial by 3000 ^{14}C yr BP; this suggests a lake level about 3-4 m below modern.

For the remainder of the Holocene, the lake level probably continued to decrease, as marked by the presence of aquatic macrophyte remains near the top of the central core. The vegetation at this time was dominated by the primary boreal forest species, but the quantity of *Picea* on the landscape was decreasing, possibly the result of cooler temperatures, increased paludification, and permafrost development.

CHAPTER 4—BIRCH LAKE

Introduction

Birch Lake has been the site of palynological studies since the early 1970s. The original pollen analysis by Ager (1975) provides a very good outline of vegetation change in central Alaska since about 14,000 ^{14}C yr BP. However, the use of conventional radiocarbon dates on bulk samples, and a uniform sample interval, means the chronology is suspect and much fine-scaled information on vegetation and climate change is missing. The present study, uses small interval sampling of selected levels, AMS dates on pollen or identifiable macrofossils, and transect cores to reconstruct lake levels. Thus providing new paleoecological information.

Birch Lake was cored as part of an on-going interdisciplinary project on lake level changes in central Alaska. My contribution to this work consist of pollen analysis from three of the shallow water cores (Birch B, D, and F) and close interval pollen analysis of one of the deep water cores (Birch H). Wendy Eisner analyzed the pollen in the remainder of the deep water cores (Birch G and H).

Location

Birch Lake (64°19'N, 146°50'W) is located in the Tanana river floodplain, nestled against the hills of the Yukon-Tanana upland (Figure 2.1). The lake basin was probably scoured by streams draining the south-side of the upland. The current lake was impounded by Tanana river glacial outwash (early Wisconsinian) which dammed streams draining the upland (Péwé and Reger, 1983). The hills (approximately 300 m to 600 m tall), are formed of highly weathered schist and are capped by wind blown silt (loess).

Boreal forest surrounds the lake, with varying proportions of *Betula papyrifera*, *Picea mariana*, *P. glauca*, *Populus tremuloides*, and *P. balsamifera*, depending on micro-site environmental conditions, especially drainage. Shoreline vegetation is dominated by *Carex* spp., and aquatic vegetation includes *Isoetes* spp., *Nuphar polysepalum*, and *Potomageton* spp.

The lake is 30 ha in area, comprises two basins, and has a maximum depth of about 13.5 m (Figure 4.1).

Results

A series of more than 10 cores were collected from Birch Lake for pollen and lake level analysis. For the purposes of this study, the discussion focuses two cores from the deepest basin, Birch G and Birch H (Figure 4.1). These two cores were combined to make a composite core (Birch G/H) that extends from the present to about 14,000 ^{14}C yr BP. Birch G and Birch H were collected within 5 m of each other, with only a 10 cm difference in water depth. To combine the cores, sample depths (from the water surface) were recalculated to a uniform water depth. The pollen stratigraphy was also compared between the two cores to confirm the correlation.

Core sediment descriptions. From the core top to 1576 cm, the core is composed of lacustrine gyttja with varying quantities of silt (Figure 4.2). Mica flecks are present throughout the core, indicating nearly continuous input of eolian silt. Inorganic silty bands are present at 1413 cm and 1520 cm. Laminations are also present throughout this section, but are mainly quite faint, with thicknesses of .5 mm or less. Three levels, however, have strong, thicker (ca. 1 mm) laminations at 1535-1538 cm, 1544-1550 cm, and 1550-1576 cm. In these levels, the laminations are composed of alternating layers of oxidized, organic, and silty sediments.

Below 1576 cm, the sediments are massive, with gradually decreasing organic content. An inorganic silty band is preserved at about 1613 cm. At 1700-1715 cm the sediments are slightly coarser and increasingly inorganic, although below this level, the sediments become slightly more organic. Beginning about 1750 cm, the sediments gradually become more inorganic, until the core base, where the sediments are composed entirely of inorganic silt.

Radiocarbon dates and age models. A series of eight AMS dates were processed from Birch G and Birch H (Table 4.1)

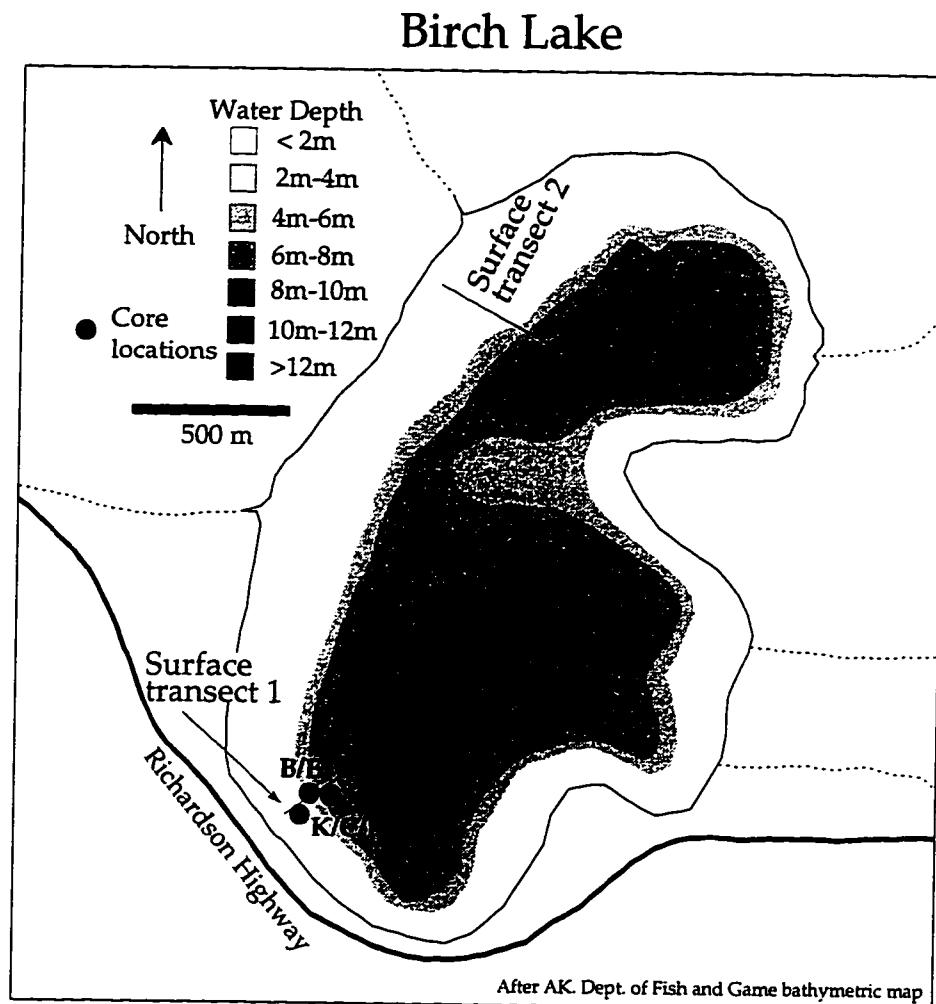


Figure 4.1: Birch Lake bathymetry.

Birch Lake Core G/H

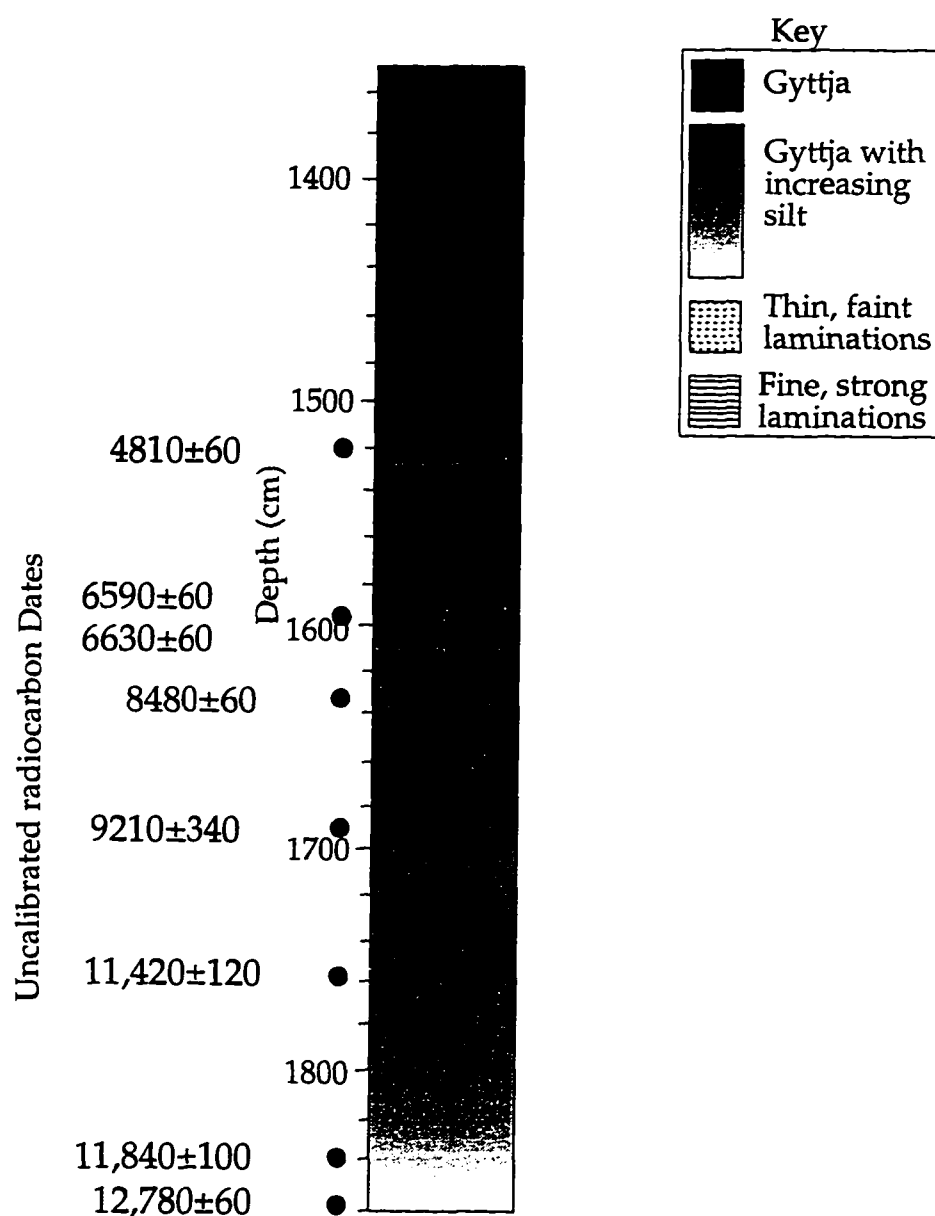


Figure 4.2: Birch Lake G/H core stratigraphy and uncalibrated radiocarbon dates.

Table 4.1: AMS dates from Birch G/H.

Depth (cm)	Material dated	mgC	Date	Lab No.	Comments
1519-1521	<i>Picea</i> pollen	0.24	4810±60	CAMS-25425	terrestrial carbon
1593-1595	<i>Picea</i> pollen	0.43	6590±60	CAMS-25426	terrestrial carbon
1593-1595	Aquatic macros	0.12	6630±90	CAMS-25422	aquatic carbon
1632-1634	<i>Picea</i> pollen	0.28	8480±60	CAMS-25427	terrestrial carbon
1690-1692	Macrofossil	0.03	9210±340	CAMS-25423	terr./aqu. carbon
1758-1759	Macrofossil	0.10	11,420±120	CAMS-25424	terr./aqu. carbon
1841-1843	Seeds	0.19	11,840±100	CAMS-25420	terrestrial carbon
1864	Seeds	1.01	12,780±60	CAMS-24521	terrestrial carbon

The Birch G/H radiocarbon dates were calibrated using CALIB 3.0, the results are presented below (Table 4.2).

Table 4.2: Calibrated ages from Birch G/H.

Uncalibrated date	Intercepts BP	Mean Intercept	1 S.D. calibrated range (BP)
4810±60	5580	5580	5474-5599°
6590±60	7400, 7430	7415	7390-7521°
6630±90	7470, 7500, 7520	7497	7391-8540°
8480±60	9450	9450	9397-9491 [‡]
9210±340	10,140; 10,270	10,205	9895-10,795†
11,420±120	13,330	13,330	13,196-13,484*
11,840±100	13,800	13,800	13,651-13,965*
12,780±60	15,080	15,080	14,910-15,270*

°Calibrated with Pearson and Stuiver (1993) bidecadal bristle cone pine calibration.

°Calibrated with Stuiver and Pearson (1993) bidecadal bristle cone pine calibration.

°Calibrated with Pearson et al., 1993 and Linick et al., 1986, bristle cone pine and German and Irish oak calibration.

† Calibrated with Kromer and Becker (1993) German oak and pine calibration.

*Calibrated with Bard et al. 1993 U-Th calibration on Barbados corals (spline fitted to oak/pine calibration).

The age model is a second order polynomial regression of the uncalibrated radiocarbon dates (Figure 4.3). A second order polynomial fits the radiocarbon dates better than a linear regression ($r^2=0.991$ and 0.977 , respectively). The model also allows for changes in sedimentation rates. The sedimentation rate in the 2nd order polynomial ranges from .06 cm/yr at the core base to .03 cm/yr at the core top. The high sedimentation rate at the core base is presumably due to the higher inorganic input there.

A second order polynomial was also chosen for the calibrated dates (Figure 4.3). The calibrated dates, aside from higher sample ages at the core base, also produce a lower sedimentation rate (.04 cm/yr to .03 cm/yr) relative to the uncalibrated age model. This is because calibrated ages become increasingly older relative to their uncalibrated counterparts. The result is a larger time interval between samples, yielding a lower sedimentation rate.

Magnetic susceptibility and Loss on ignition (LOI). Magnetic susceptibility and loss-on-ignition analyses were conducted by Bruce Finney, and are presented here with his permission.

Magnetic susceptibility is high at the core base (about 12,500 ^{14}C yr BP), but rapidly decreases thereafter, until about 6000 ^{14}C yr BP (after which it remains unchanged until the core top). Episodes of higher susceptibility at about 11,000 ^{14}C yr BP, 10,000 ^{14}C yr BP, and 7400 ^{14}C yr BP interrupt the overall decrease in susceptibility (Figure 4.4). No tephtras were recognized in the core, and the higher susceptibility probably reflects periods of increased sediment flux into the Birch Lake basin.

LOI (500°C) is low from about 12,500 to about 9600 ^{14}C yr BP. Between 9600 and 8200 ^{14}C yr BP, LOI increases, and subsequently decreases between 8000 and 7000 ^{14}C yr BP. LOI then increases, reaching maximum values about 5600 ^{14}C yr BP. Subsequently, LOI decreases slightly to modern levels (Figure 4.4).

The curve for Loss-on-ignition at 850° C shows two levels of carbonate enrichment about 10,700 and 8800 ^{14}C yr BP (Figure 4.4). Close examination of the core suggests ostracod shells are the reason for the higher carbonate content.

Pollen zones. Pollen zones were identified both visually and through computer-generated clustering algorithms. Comparison between the computer-generated zones of the actual and randomized data of Birch G/H indicate 11 (CONISS), nine (SPLITINF), or eight (SPLITLSQ) significant zones (Figure 4.5). All the computer algorithms identified boundaries at about 12,200 ^{14}C yr BP (*Betula* rise-1837 cm), 11,000 ^{14}C yr BP (increasing *Artemisia*-1775 cm), 9300 ^{14}C yr BP (*Populus* rise-1690 cm), 8000 ^{14}C yr BP (*Picea* rise-1637 cm), 7000 ^{14}C yr BP (*Alnus* rise-1590 cm), and 5300 ^{14}C yr BP (increasing *Picea* %-1532 cm) (Figure 4.6). The zone boundaries used for this analysis are the main boundaries (*Betula*, *Picea*, and *Alnus* rises), in addition to the *Populus*

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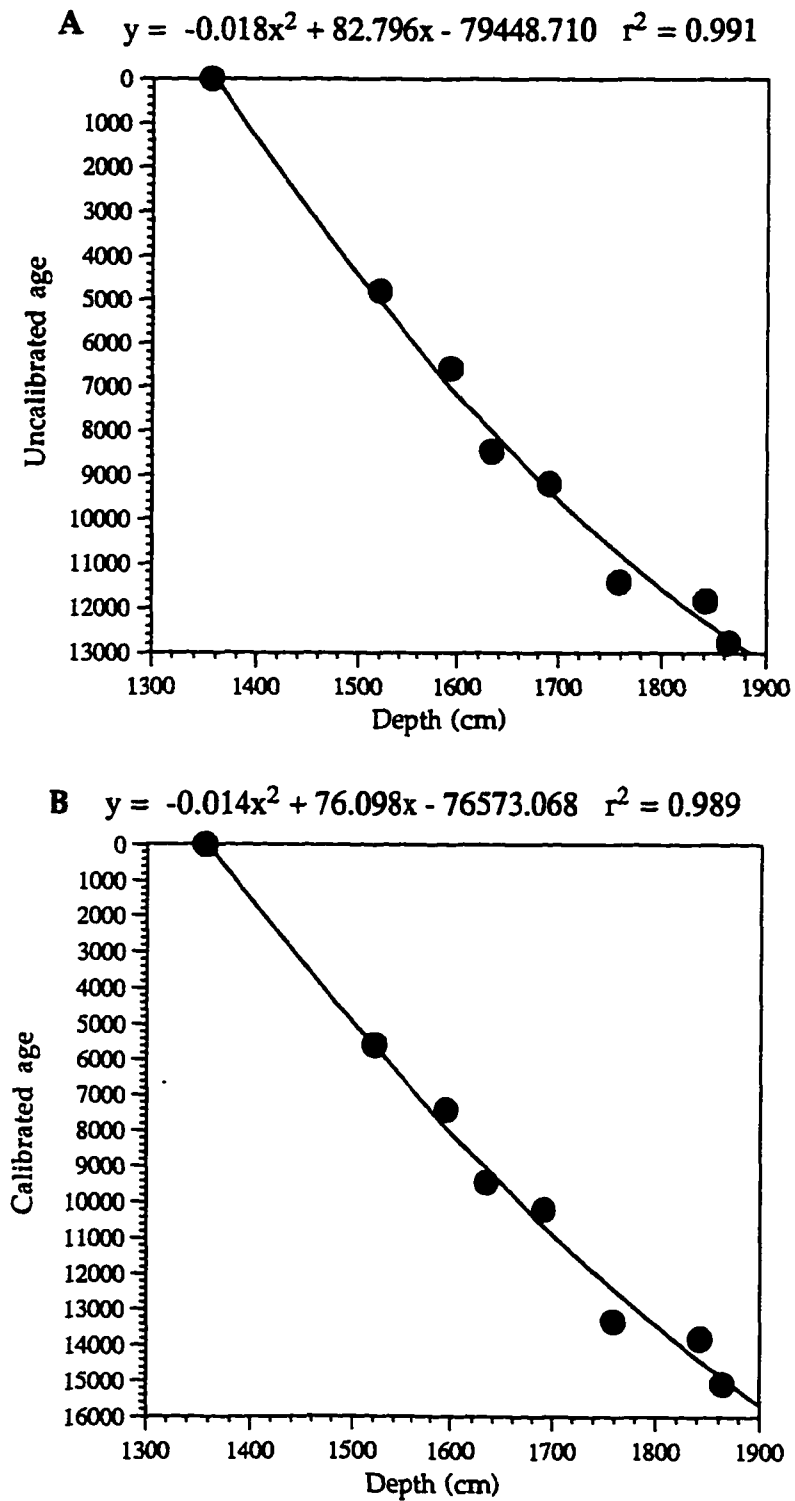


Figure 4.3: Uncalibrated (A) and calibrated (B) age models for Birch Lake.

Birch Lake Core G/H

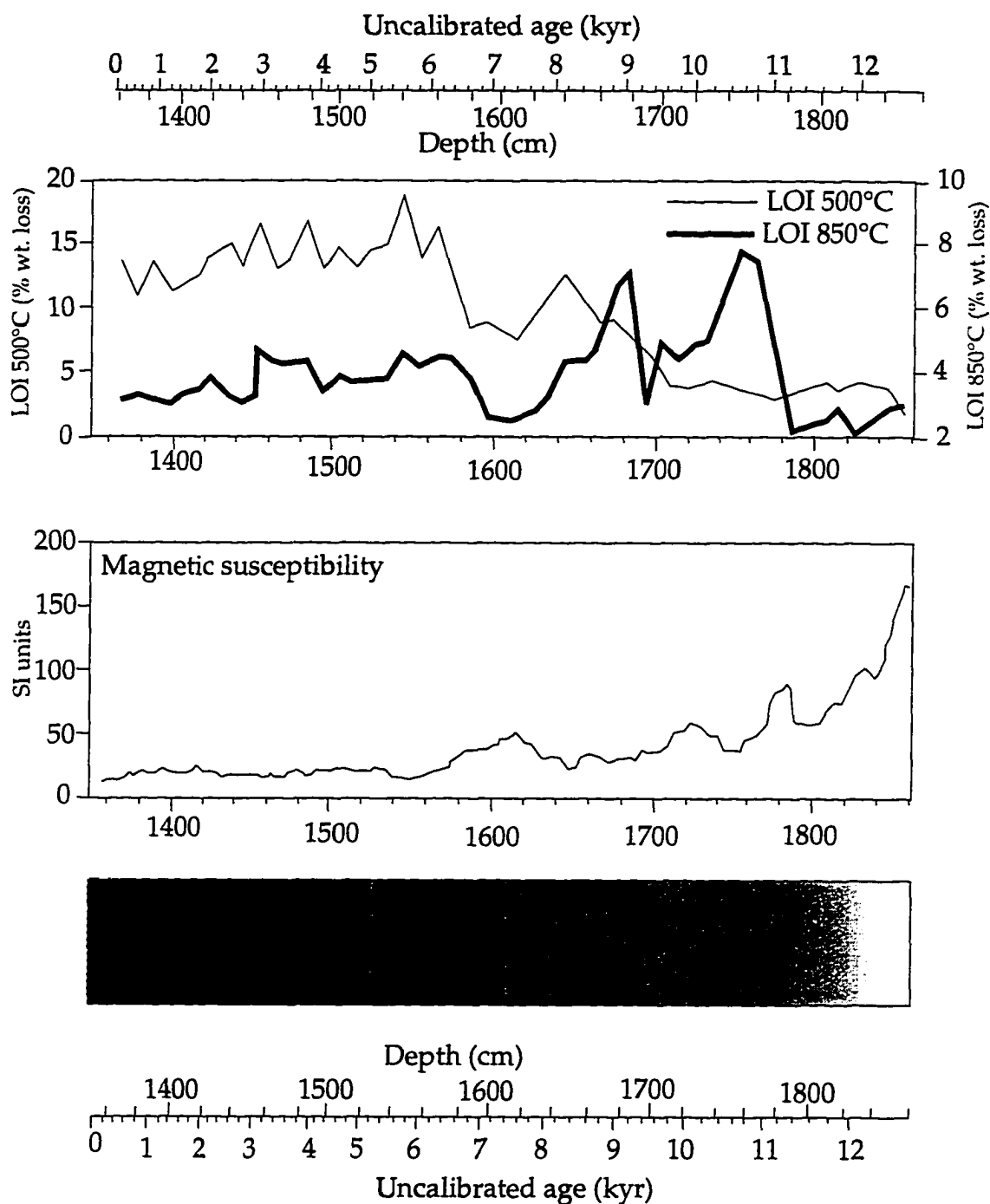


Figure 4.4: Birch Lake core G/H stratigraphy, LOI, and magnetic susceptibility. See Figure 4.2 for a key to stratigraphic patterns.

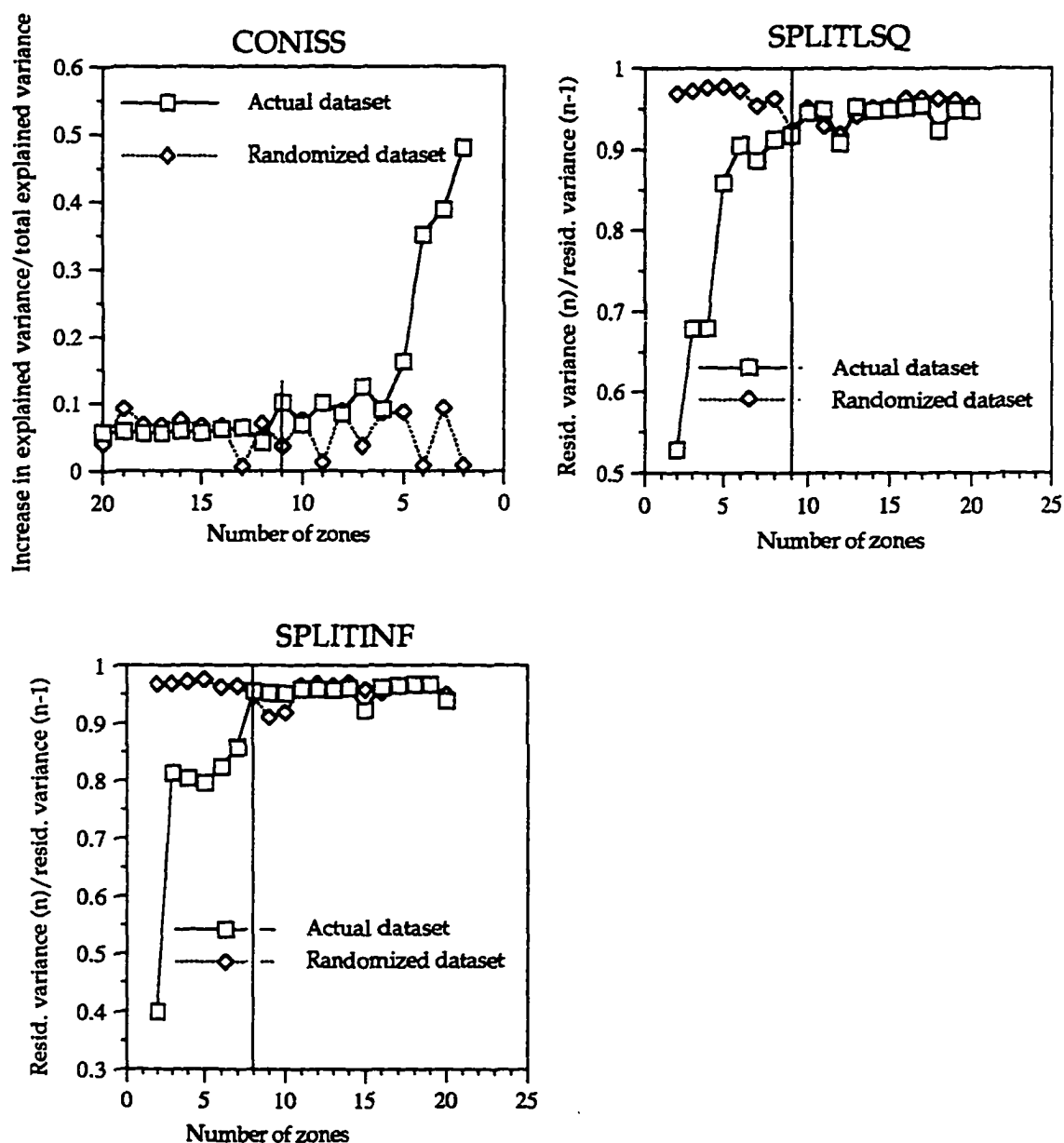


Figure 4.5: Birch core G/H, zoning of actual and randomized datasets, lines mark where datasets converge or diverge.

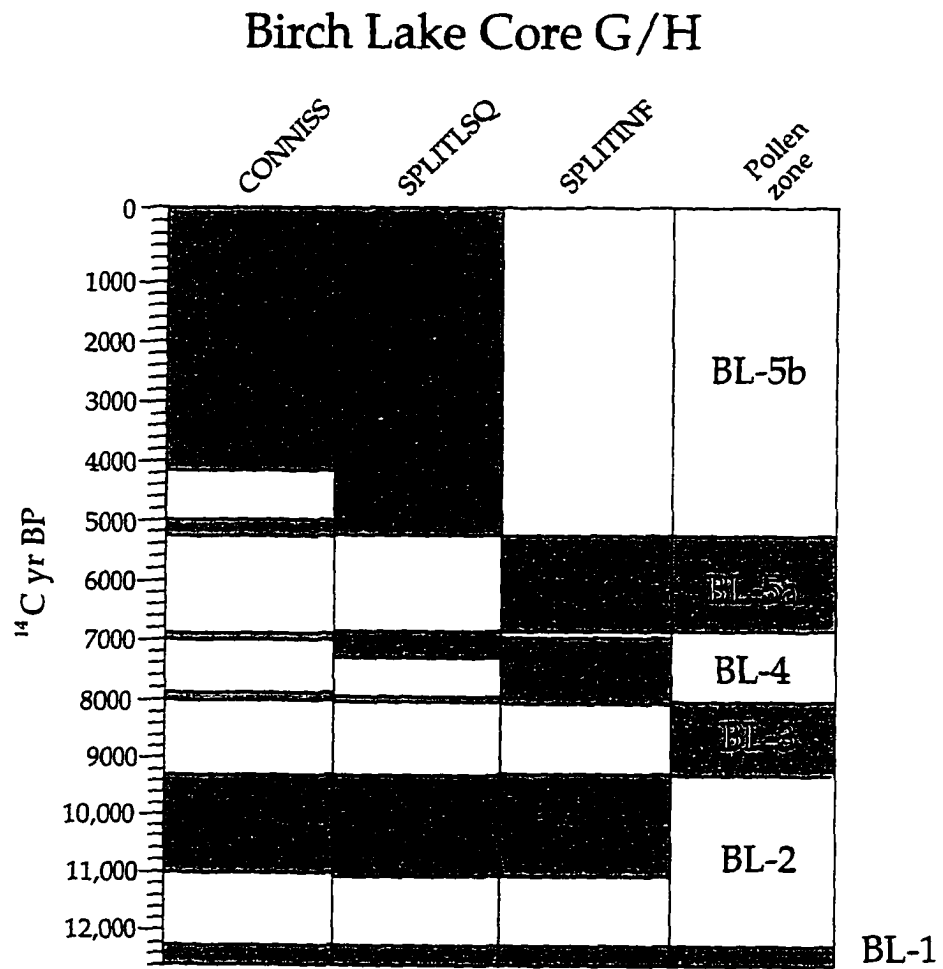


Figure 4.6: Birch Lake core G/H pollen zones.

rise. The boundary at 5300 ^{14}C yr BP marks a subzone boundary because it reflects relative changes in the dominant boreal forest taxa. The boundary at 11,000 ^{14}C yr BP was not used because it marks a less significant vegetation change than the other boundaries. In addition, the sample interval decreases at this boundary, indicating that the zoning algorithms may be sensitive to variations in sample interval, irrespective of whether the vegetation changes.

Pollen zone BL-1 extends from the core base (about 12,500 ^{14}C yr BP) to 12,200 ^{14}C yr BP (1860 to 1837 cm) (Figure 4.7). While this zone only consists of three samples, these samples are consistent with herb zone samples from earlier Birch Lake cores (Ager, 1975). That is, they are characterized by low *Betula*, and high *Poaceae*, *Cyperaceae*, and *Artemisia* percentages. Influx analysis indicates low *Betula* accumulation, with *Salix*, *Artemisia*, and *Cyperaceae* dominating the shrub/herb layer (Figure 4.8). *Pediastrum*, a green alga is very common in this zone.

Pollen zone BL-2 extends from about 12,200 to 9300 ^{14}C yr BP (1837 to 1690 cm). This zone is marked by the rapid increase and dominance of *Betula*, along with decreases in *Salix*, *Artemisia*, and *Poaceae* percentages (Figure 4.7). *Cyperaceae* remain at their previous levels. Influx analysis indicates *Betula*, *Salix*, and *Poaceae* remain relatively unchanged, *Cyperaceae* increase, and other herbs decrease in this zone (Figure 4.8).

High resolution (1-2 cm interval) pollen analysis in the middle of this zone (11,000 to 10,100 ^{14}C yr BP [1775-1730.5 cm]) indicates several short-term changes in the pollen assemblage. *Juniperus*, *Salix*, *Artemisia*, and *Poaceae* percentages increase slightly, while *Betula* and *Cyperaceae* decrease or remain unchanged. *Pediastrum* nets are also common in this interval. Influx analysis indicates that *Juniperus*, and *Artemisia* increase, while *Poaceae* and *Cyperaceae* remained unchanged. These data suggest that although the landscape was still *Betula*-dominated, there was an expansion of xerophytic taxa. The relative changes in the pollen assemblage in this interval are very small, but similarities with Windmill Lake suggest the changes reflect actual transformations on the landscape. This is discussed further in Chapter 5.

Pollen zone BL-3 dates between about 9300 and 8100 ^{14}C yr BP (1690 to 1640 cm). This zone is characterized by continued high *Betula* percentages, but with the addition of *Populus*, which comprises almost 15% of the pollen sum

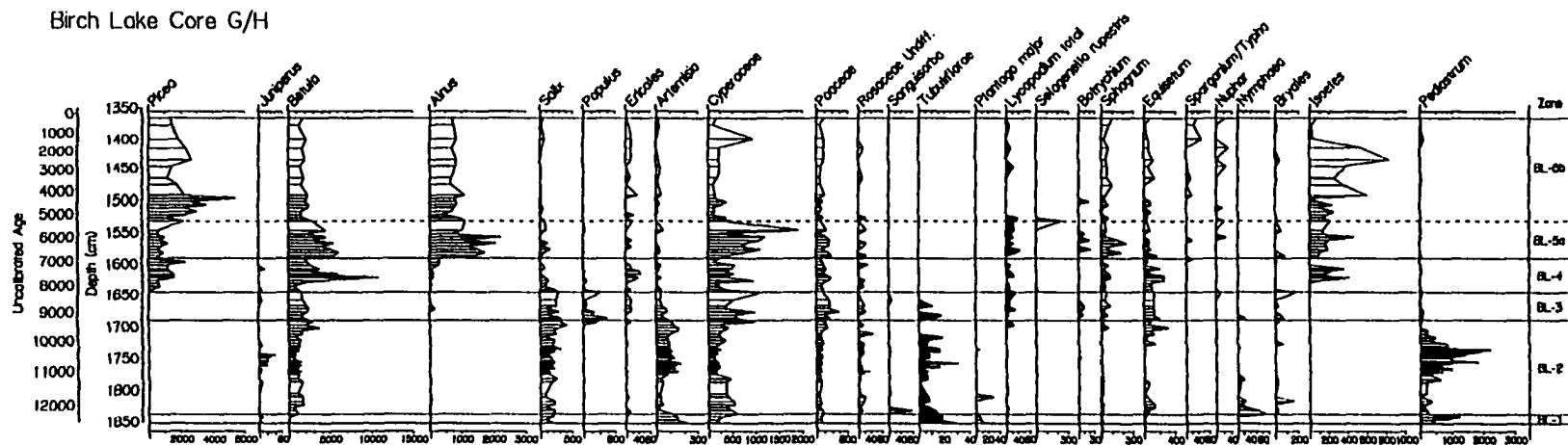


Figure 4.8: Birch G/H pollen influx (grains/cm²/yr).

(Figure 4.7). Influx analysis indicates continued *Betula* and *Salix*-dominated landscape (Figure 4.8), although the marked increase in *Populus* suggests this taxa may have occupied sites surrounding Birch Lake, or on the Tanana River floodplain.

Pollen zone BL-4 dates between about 8100 and 6900 ^{14}C yr BP (1640 to 1590 cm). This zone is characterized by the appearance of *Picea* and the continued dominance of *Betula* (Figure 4.7). *Populus* is virtually absent. Influx analysis indicates that *Betula* accumulation briefly peaked with the *Picea* arrival, but then returned to lower levels (Figure 4.8). *Salix* and *Artemisia* influx decrease during this zone, although Cyperaceae and Poaceae are relatively unchanged.

Pollen zone BL-5 dates between 6900 ^{14}C yr BP and the present (1590 cm to the core top) (Figures 4.7 and 4.8). This zone is characterized by the arrival of *Alnus*, and the co-dominance of *Picea*, *Betula*, and *Alnus* in the pollen assemblage.

The transition between subzone BL-5a to BL-5b occurs about 5300 ^{14}C yr BP (1532 cm). Subzone BL-5a is characterized by moderate *Picea* frequencies, high, but declining *Betula*, and high *Alnus* frequencies. Subzone BL-5b is marked by increasing *Picea* and decreasing *Betula* frequencies. Influx analysis indicates that *Picea* accumulation is at minimum levels in subzone BL-5a, but rapidly increases to maximum values in subzone BL-5b, and subsequently decreases. *Betula* influx shows a small peak in the beginning of subzone BL-5a, but subsequently decreases and is stable at the onset of subzone BL-5b. *Alnus* influx is at maximum values in subzone BL-5a and subsequently decreases to modern levels. Cyperaceae influx is at maximum levels in subzone BL-5a, decreasing to modern values in subzone BL-5b.

Rate of Change. Rate of change analysis is used to identify periods where the proportion of pollen taxa changes rapidly in adjacent samples (Bennett, 1996).

Rate of change analysis was conducted on the pollen percentage data, using the taxa listed below (Table 4.3).

Table 4.3: Taxa used in rate of change analysis.

<i>Picea</i> , undifferentiated
<i>Juniperus</i>
<i>Betula</i>
<i>Alnus</i>
<i>Salix</i>
<i>Populus</i>
<i>Ericales</i>
Rosaceae
<i>Sanguisorba</i>
<i>Solidago</i> -type
<i>Artemisia</i>
Cyperaceae
Gramineae

Chord distance is the dissimilarity coefficient (DC) used for this analysis, because it has a good signal to noise ratio (Jacobsen et al., 1987). To remove the effect of varying sample intervals, the rate of change (DC/100 ^{14}C yr) was calculated on an interpolated dataset with a 100-year sample interval.

Unsurprisingly, the highest rate of change occurs near the Birch G/H base, at the transition from pollen zone BL-1 to BL-2, about 12,200 ^{14}C yr BP (Figure 4.9). Subsequently, between about 11,000 and 4000 ^{14}C yr BP, there are several peaks in rate of change, the highest occur when *Picea* and *Alnus* pollen appear in the core. The upper part of the core shows very little change, but this is due to the coarse sample interval there.

Ordination. Ordination of the taxa listed in Table 4.3 yielded structure in the dataset. The gradient length for the first axis on square root-transformed percentage data is 1.833 standard deviation units, indicating Principle Components Analysis (PCA) is an appropriate method for the ordination. PCA axes 1 and 2 explain 69.0% and 11.6%, respectively, of the variation. The PCA biplot of the sample scores (Figure 4.10) separates the data into four main groups. Pollen zone BL-1 is marked by negative loadings on axis 1, and positive loadings on axis 2. Pollen zones BL-2 and BL-3 are tightly clustered together with axis 1 loadings near zero, and negative axis 2 loadings. Pollen zones BL-4 and BL-5a form a somewhat diffuse cluster with slightly positive loadings on axis 1 and slightly negative loadings on axis 2. Finally subzone BL-5b is marked by positive axis 1 loadings and axis 2 loadings that are near zero.

Birch Lake Core G/H

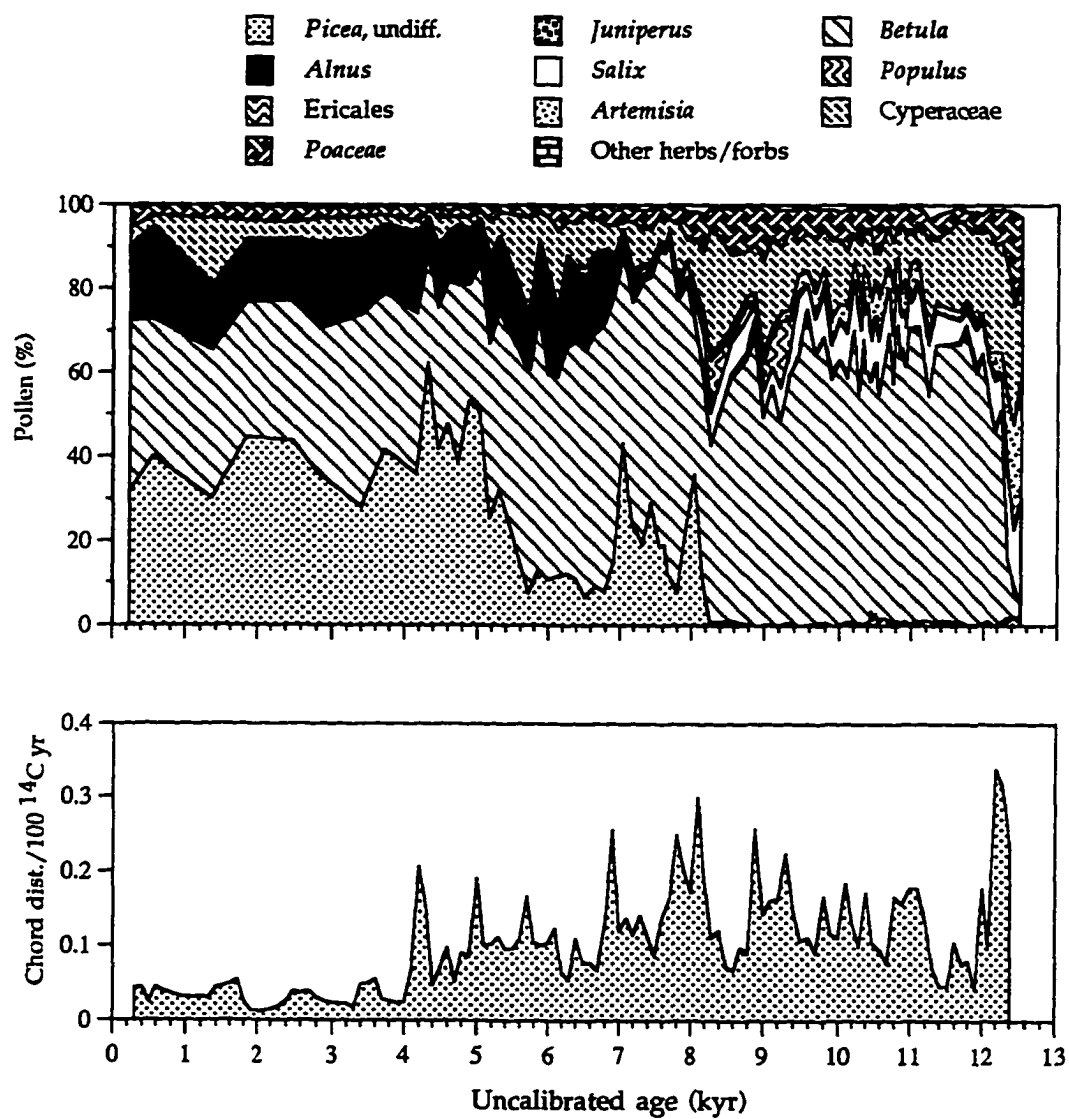
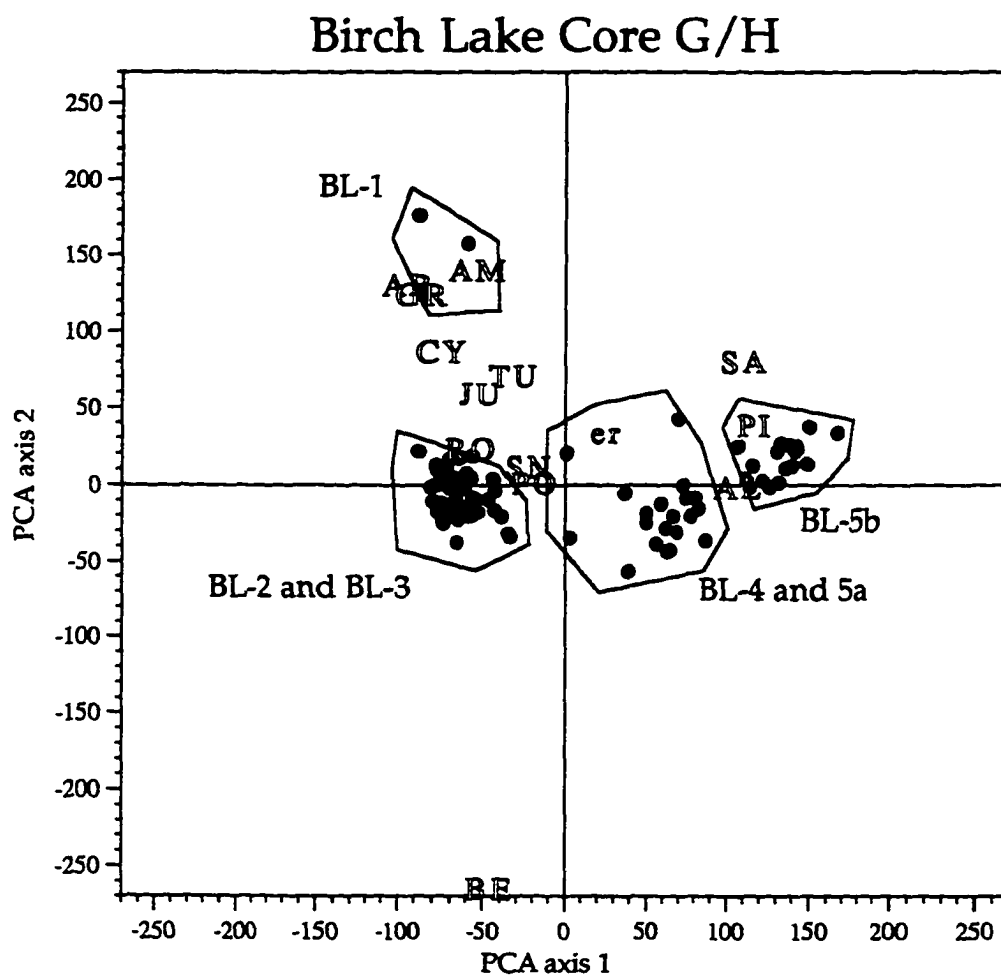


Figure 4.9: Birch core G/H rate of change, uncalibrated age scale.



Taxon abbreviations

PI= <i>Picea</i> undiff.	AM= <i>Solidago</i> -type
BE= <i>Betula</i>	TU= <i>Tubulifloreae</i> undiff.
AL= <i>Alnus</i>	SN= <i>Sanguisorba</i>
er= <i>Ericales</i> undiff	RO= <i>Rosaceae</i>
PO= <i>Populus</i>	
JU= <i>Juniperus</i>	
CY= <i>Cyperaceae</i>	
SA= <i>Salix</i>	
GR= <i>Poaceae</i>	
AR= <i>Artemisia</i>	

Figure 4.10: Birch G/H PCA biplot.

When the PCA sample scores are plotted against time, additional patterns emerge (Figure 4.11). Axis 1 is a monotonic trend from low to high scores; it is highly correlated with *Picea* and *Alnus* abundances. Axis 2 for much of the core varies close to zero, indicating that little of the pollen variability is modeled on this axis. One exception, however, is at the core base, where axis 2 values are highly positive.

The species scores indicate that *Picea*, and *Alnus*, have high positive loadings on axis 1, while *Betula*, *Juniperus*, *Salix*, and the herbaceous taxa have high negative loadings. With axis 2, *Betula* has high negative loadings, while the herbaceous taxa have strong positive loadings (Table 4.4).

Table 4.4: PCA species loadings.

Species	Axis 1	Axis 2
<i>Picea</i> undiff.	117	36
<i>Juniperus</i>	-51	57
<i>Betula</i>	-46	-267
<i>Alnus</i>	106	-4
<i>Salix</i>	-110	78
<i>Populus</i>	-19	-1
Ericales	25	32
<i>Artemisia</i>	-95	128
Cyperaceae	-75	85
Poaceae	-86	122
Other herbs (mean)	-42	59

C and N concentrations and isotopes. Measurements of C and N isotopes and concentrations were conducted by Bruce Finney, and are used here with his permission.

C and N concentrations were measured on total sedimentary organics. At the base of Birch G/H, C and N concentrations are low, but increase dramatically at about 9400 ^{14}C yr BP (Figure 4.12). They subsequently decrease between *ca.* 8000 and 6500 ^{14}C yr BP, and then increase to near modern levels by 5700 ^{14}C yr BP, remaining relatively stable until the core top.

Carbon and Nitrogen fluxes ($\text{g}/\text{cm}^2/\text{yr}$) are high from about 12,400 ^{14}C yr BP to about 9600 ^{14}C yr BP (Figure 4.12), mainly because of high bulk density and sedimentation rates in this part of the core. C and N fluxes increase markedly about 9600 ^{14}C yr BP, this coincides with higher C and N

Birch Lake Core G/H

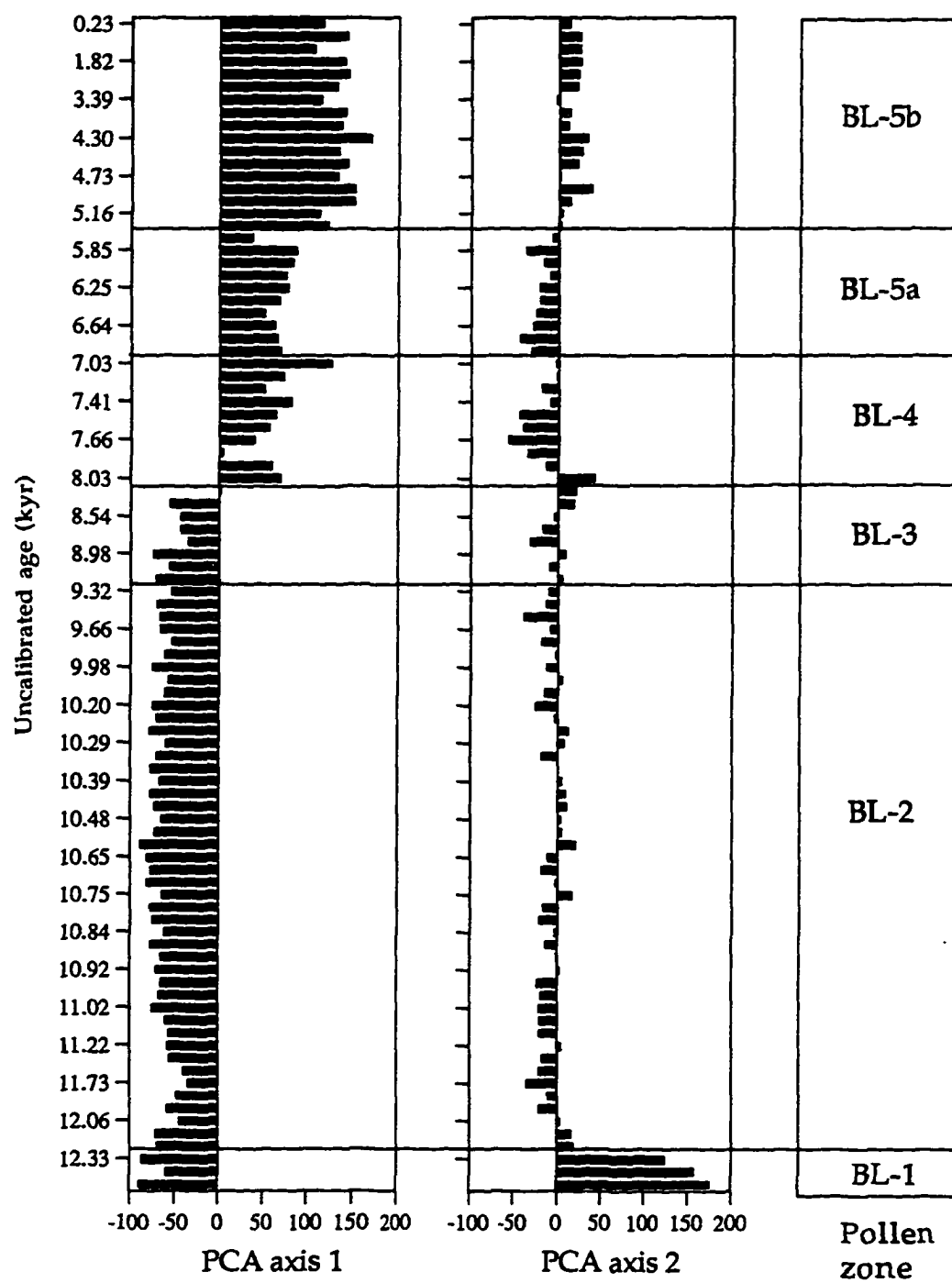


Figure 4.11: Birch Lake core G/H PCA scores plotted against age.

Birch Lake Core G/H

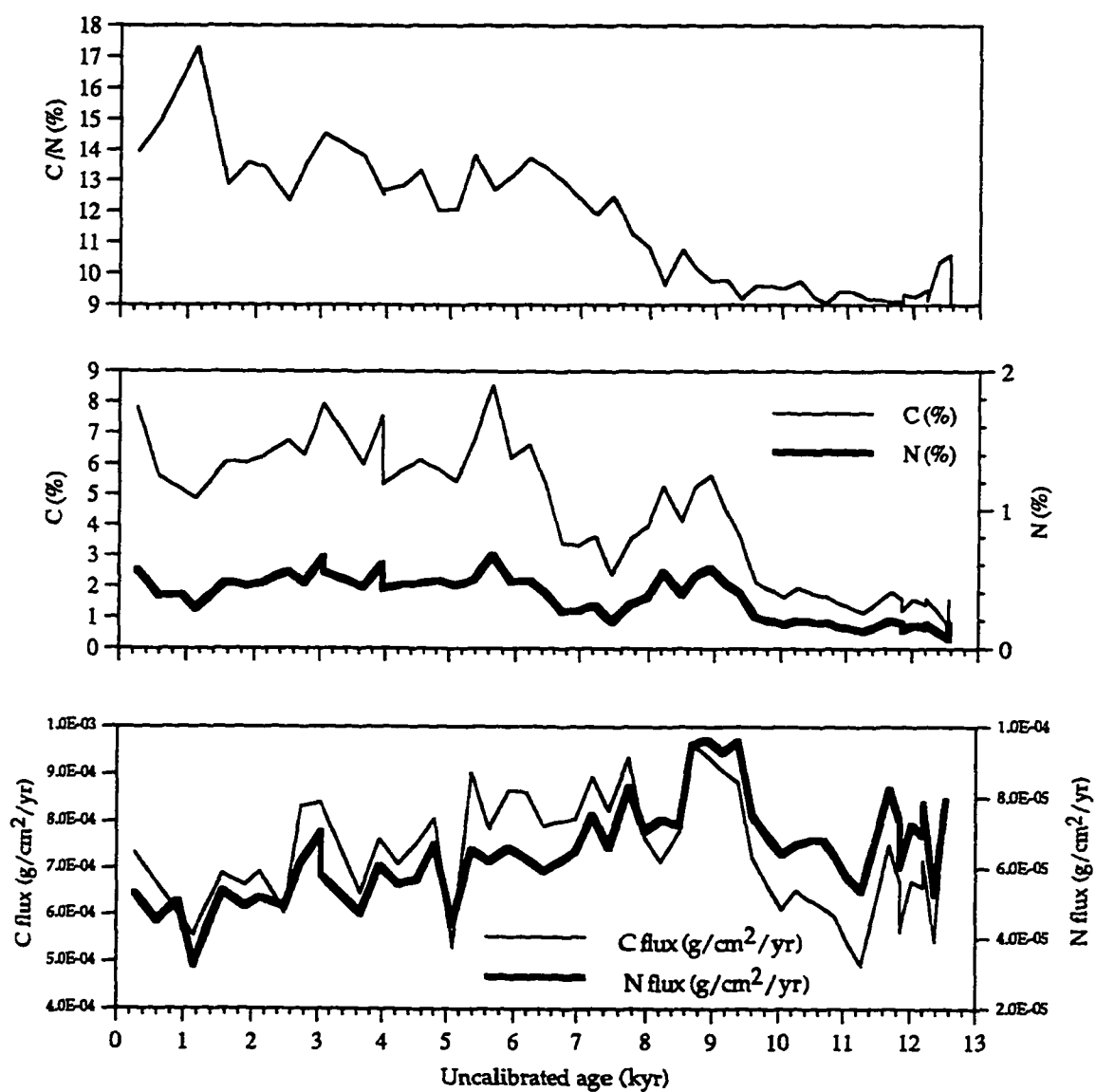


Figure 4.12: Birch Lake core G/H, carbon and nitrogen concentrations, C/N ratios, and fluxes.

concentrations, indicating increasing lake productivity. After about 9000 ^{14}C yr BP, C and N fluxes decrease gradually, probably reflecting the lower sedimentation rate, as C and N concentrations remain high (Figure 4.12).

Carbon and Nitrogen ratios are low (<10) in the lower part of the core; they start increasing at about 9000 ^{14}C yr BP and achieve modern levels (about 13) by about 6200 ^{14}C yr BP (Figure 4.12).

$\delta^{13}\text{C}$ values are high at the core base, decrease rapidly at 12,300 ^{14}C yr BP, and increase at about 11,000 ^{14}C yr BP. From 11,000 to 9600 ^{14}C yr BP, $\delta^{13}\text{C}$ is high, but decreases markedly at 9400 ^{14}C yr BP, reaching a core low 8900 ^{14}C yr BP. Subsequently, $\delta^{13}\text{C}$ increases and reaches a core high at about 7500 ^{14}C yr BP and then gradually decreases to modern levels (Figure 4.13).

$\delta^{15}\text{N}$ is slightly elevated at the core base, but subsequently decreases at about 11,600 ^{14}C yr BP (Figure 4.13). From about 11,600 ^{14}C yr BP to about 8900 ^{14}C yr BP, $\delta^{15}\text{N}$ gradually increases, achieving modern levels at 8600 ^{14}C yr BP. $\delta^{15}\text{N}$ varies considerably in the top four samples in the core. Single samples mark these excursions, suggesting these points reflect analytical error.

Discussion

Vegetation reconstruction. The earliest vegetation documented at Birch Lake is a sparse herbaceous and *Salix* assemblage (Figures 4.7 and 4.8). While only a few herb zone samples are preserved in the Birch G/H core, similarities with an earlier core collected by Ager (1975), indicate the landscape was sparsely covered with *Salix*, grasses, sedges, and *Artemisia*. The transition to a shrub *Betula*-dominated vegetation was abrupt and occurred later here than previously thought. The original date on the birch rise from Birch lake was about 14,700 ^{14}C yr BP (Ager, 1975). The new AMS date from Birch G/H suggest this event took place significantly later, about 11,800 ^{14}C yr BP (although the age models suggests it occurred at 12,300 ^{14}C yr BP). Influx analysis indicates that while *Betula* was an important species, total pollen influx for all terrestrial taxa was lower than modern values, suggesting the vegetation may have been sparser than it is today (Figure 4.8). Although some of this difference may be due to the absence of abundant pollen producers, such as *Picea* and *Alnus*. The influx of herbaceous taxa is

Birch Lake Core G/H

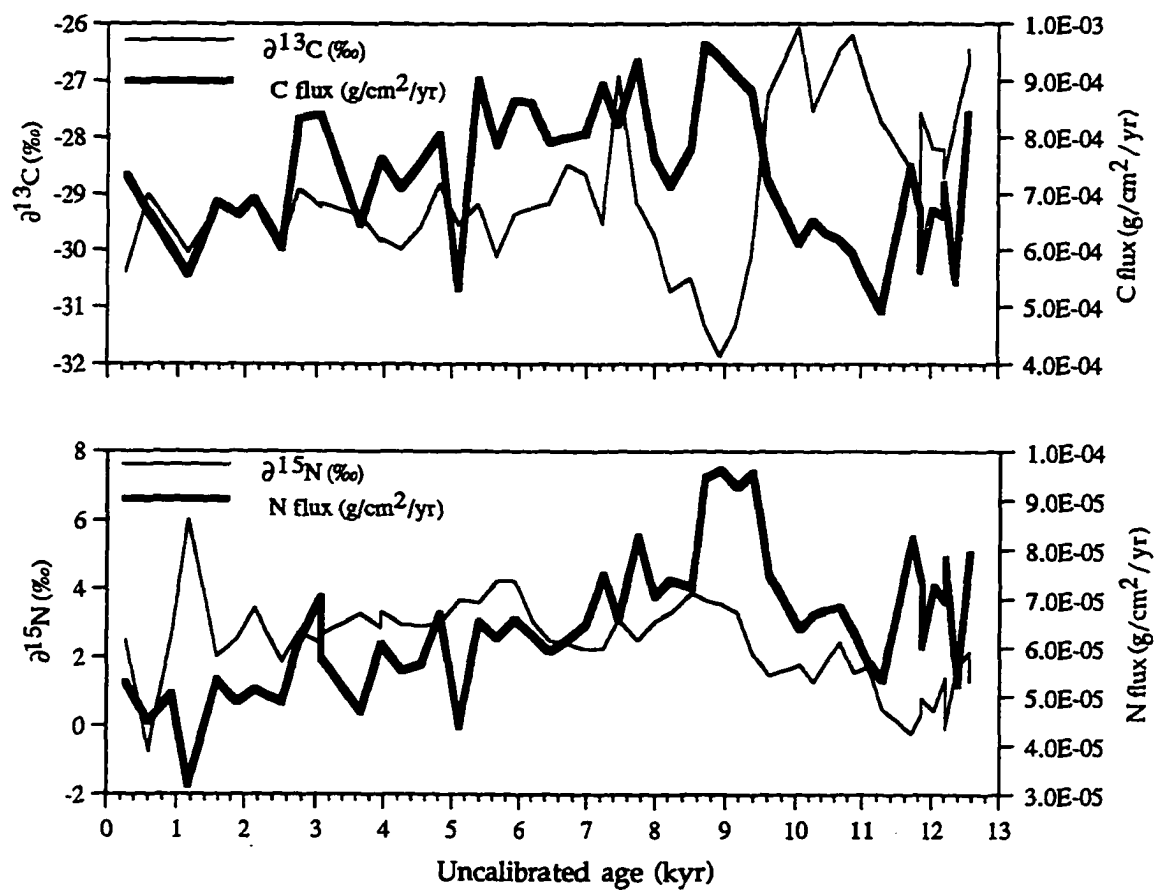


Figure 4.13: Birch Lake core G/H carbon and nitrogen isotopes and fluxes.

unchanged during the transition, indicating they were as common during the birch zone as during the herb zone. These patterns of rapid *Betula* expansion but continued abundant herbaceous ground cover have also been documented at Windmill in addition to several other central Alaskan lakes (Hu et al., 1993; Edwards and Brubaker, 1986; Anderson et al., 1988).

The rapid, nearly instantaneous expansion of *Betula* suggests the species was present during the herb zone in low densities in scattered locations. *Betula* pollen has been recovered from Bering sea cores dating to this period, suggesting it grew and sexually reproduced on some parts of the Bering Land Bridge (Elias et al., 1996). The expansion of *Betula* was probably triggered by combined higher summer temperature and precipitation. Summer solar insolation was rapidly increasing at this time (Berger, 1978), and lake level analyses at both Birch and Windmill suggest lake levels were increasing about 12,000 ^{14}C yr BP, indicating higher effective moisture.

Between roughly 11,000 and 10,000 ^{14}C yr BP, slightly elevated *Artemisia* frequencies from Birch Lake suggests a change in the vegetation (Figure 4.7). Influx analysis suggests *Artemisia* and *Juniperus* had higher abundances, while *Cyperaceae* abundance decreased (Figure 4.8). While these changes are small, they reflect a real alteration of the landscape.

In central Alaska, *Artemisia frigida* and *Juniperus communis* prefer dry, sunny locations that have a thin winter snow pack. Expansion of these taxa around Birch and Windmill lakes probably occurred on ground surfaces susceptible to increased aridity, such as south-facing slopes or areas with good drainage, such as active flood plains. Why the vegetation change is apparently less at Birch than Windmill may be due to different pollen catchment properties of the lakes, the different elevations of the lakes, or to regional differences in climate. This is discussed further in Chapter 5.

In addition, as at Windmill Lake, Birch lake also has very high *Pediastrum* accumulation at this time. The significance of the *Pediastrum* is not entirely clear, but many *Pediastrum* species prefer lakes with high conductivity and pH (Crisman, 1978), suggesting that the *Pediastrum* reflects episodes of increased sediment and nutrient flux into Birch lake. Today, primary air-fall loess can be rich in carbonates and other cations (Rieger, et al., 1979); increased loess deposition during the late-glacial may have resulted in *Pediastrum*

blooms at Birch and Windmill Lakes, although other factors, such as summer temperature and lake productivity may have also been important.

Between about 9200 and 8200 ^{14}C yr BP, *Populus* pollen at Birch Lake increases from near zero to about 15% of the terrestrial vegetation (Figure 4.7). *Populus* pollen does not preserve well, nor does it travel far from its source (Sangster and Dale, 1961; Edwards and Dunwiddie, 1985), indicating it marks a significant change in the local, rather than regional, vegetation.

The *Populus* at this time period may be *P. balsamifera* because this taxon was identified (based on slight differences in the pollen grains) in a lacustrine record in the central Brooks Range (Brubaker et al., 1983).

P. balsamifera is a shade-intolerant pioneering species, commonly found on well-drained sites (Viereck, et al., 1992), and shows a positive growth response to warm early summer temperatures (Lev, 1987). This suggests that the expansion of *Populus* at Birch Lake and throughout central Alaska is linked not only with summer temperature, but also availability of habitat for an early successional species. To what extent either of these factors plays a role is probably unique to each locality. Birch, Windmill, and Dune lakes all have *Populus* zones that differ slightly either in timing, or in quantity of *Populus* pollen, suggesting that local factors can be very important.

For example, at Dune Lake, the *Populus* subzone is longer lived (ca. 1200 ^{14}C yr) and starts earlier (before 9700 ^{14}C yr BP) than at Birch Lake (Figure 3.12). This difference may be explained by the extensive dune field that would have provided excellent habitat for the *Populus* (which may have been *P. tremuloides*) at Dune Lake. In contrast, at Windmill Lake, the *Populus* subzone is not especially long-lived, and the *Populus* frequencies are low, less than 10% (Figure 2.8). The hillsides surrounding Windmill Lake are mostly north and east-facing; it is possible the *Populus* were limited by cool/wet soils there. At Birch Lake, the *Populus* subzone is well developed, with frequencies as great as 15%, suggesting that the trees were relatively common. They may have grown on the lakeshore, on the hillsides surrounding the lake, and/or on the Tanana river floodplain.

About 8000 ^{14}C yr BP, *Picea* appeared on the landscape around Birch Lake, resulting in a mixed vegetation dominated by *Picea* and *Betula* (Figure 4.7). Influx analysis indicates that *Betula* accumulation was at maximal levels

during the initial appearance of *Picea* (Figure 4.8). The high *Betula* accumulation may be due to the appearance of tree birch in the region. *Betula* seeds have been analyzed from only one interior lake (Wein Lake), and results suggested that birch trees were not present there until the arrival of spruce, about 9700 ^{14}C yr BP or later (Hu et al., 1993).

The major factors influencing the spruce rise were a combination of increased temperature and moisture. Higher summer temperatures are suggested by continued high summer insolation due to Milankovitch cycles, and higher lake levels at Dune and Birch (discussed below) indicate increased effective moisture at this time.

Subsequently, about 6800 ^{14}C yr BP, higher than modern influx values indicate *Alnus* became an important component of the vegetation (Figure 4.8). At both Windmill and Birch, *Alnus* first appears in the cores about the same time as spruce, or shortly before, although it does not become important until significantly later. This early *Alnus* "tail" has been documented in several central Alaskan sites (Ten Mile [Anderson et al., 1994], Farewell [Hu et al., 1996], Wonder Lake [Anderson et al., 1994], Screaming Yellowlegs [Edwards et al., 1985]), and in the western Brooks Range (Niliq [Anderson, 1988]). There does not seem to be any temporal or geographic patterns in the presence or absence of the tail. It does suggest, however, that *Alnus* was probably present in the region in the early Holocene, although in low densities.

Pollen over-represents actual *Alnus* distributions. If the 20% pollen isopoll is an accurate indicator for local presence (Anderson and Brubaker, 1986), the values at Birch suggests that *Alnus* did not dominate the landscape. This is similar to Dune, where *Alnus* is 20% to 30% of the terrestrial pollen throughout the record (Figure 3.12). *Alnus* is slightly higher at Windmill, reaching as high as 40% (Figure 2.8). These data are typical of central Alaskan sites, where *Alnus* does not dominate the vegetation. In southwest Alaska and the western Brooks Range however, *Alnus* percentages can be as high as 60% to 80% (Hu et al., 1995; Anderson, 1985). These differences are probably due to the higher effective moisture at the coastal sites.

Beginning 5200 ^{14}C yr BP, *Picea* percentages and influx reach higher than modern levels, with peak values at about 4200 ^{14}C yr BP and a second peak at

about 2600 ^{14}C yr BP (Figures 4.7 and 4.8). The significance of these changes are not clear, but higher than modern *Picea* abundances at Dune and Windmill lakes suggest that spruce was more common in the past than it is today. This topic is discussed more fully in Chapter 3, but the decreasing *Picea* abundance in the late Holocene may indicate lowered summer temperatures, increased permafrost, and paludification. This suggests that the boreal forest we see today only developed after 4000 ^{14}C yr BP, and possibly as recently as 2000 ^{14}C yr BP.

Stable carbon isotopes. Much of the particulate organic matter in the Birch G/H core probably originated within the lake itself. Low C/N ratios, as well as $\delta^{15}\text{N}$ values >0 suggest little terrestrial material has been reworked into the lake basin. This suggests that the stable carbon isotopes reflect both the $\delta^{13}\text{C}$ signal of the source carbon (in this case, the DIC of the lake water), as well as changes in the carbon prior to sedimentation (see also Chapter 2 and Appendix B for more detailed discussions).

The $\delta^{13}\text{C}$ signal of the lake water DIC is affected by the $\delta^{13}\text{C}$ of the DIC entering the lake from various sources, such as precipitation, surficial run-off, or groundwater. At Birch Lake, run-off and groundwater flux account for 50% to 66% of the water budget (B. Finney, pers. comm.), indicating that the lake water DIC is strongly controlled by these DIC sources.

About 13,000 ^{14}C yr BP $\delta^{13}\text{C}$ is high, but then decreases 12,500-11,500 ^{14}C yr BP, subsequently increases to maximum levels 11,000-9500 ^{14}C yr BP, and then decreases to minimum levels 9500 to 8500 ^{14}C yr BP (Figure 4.13). C and N concentrations as well as C/N ratios are low from about 13,000 to 9500 ^{14}C yr BP (Figure 4.12), indicating low lake productivity and probably few aquatic macrophytes. The variable $\delta^{13}\text{C}$ in this part of the core probably reflects changes in the lake water DIC. The high $\delta^{13}\text{C}$ at the core base coincides with stratigraphic changes indicating a lowered lake level, while the slightly reduced $\delta^{13}\text{C}$ from about 12,500 to 11,500 coincides with stratigraphic evidence of higher water level (see below). The high $\delta^{13}\text{C}$ from 11,000 to 9600 only somewhat coincides with stratigraphic evidence of lower water level, although the dates for this lowstand are problematic (see below).

From 9600 to 9000 ^{14}C yr BP, $\delta^{13}\text{C}$ decreases substantially. At this same time, C and N concentrations and fluxes increase, while C/N ratios continue

at low levels. The higher C and N concentrations and fluxes indicate that lake productivity was probably increasing, although not enough to affect the carbon isotope signal. The abrupt decrease in $\delta^{13}\text{C}$ probably reflects higher groundwater input in the lake's DIC, although lake levels may have actually dropped slightly between 9800 and 8800 ^{14}C yr BP (see below). After about 8500 ^{14}C yr BP, the $\delta^{13}\text{C}$ of the sediments increases, and then stabilizes at about -30 to -29‰ (Figure 4.13). C and N concentrations and fluxes increase at this time (Figure 4.12), indicating higher lake productivity. It is probable that the changes in $\delta^{13}\text{C}$ from 8500 ^{14}C yr BP to the present reflect mainly changes in lake productivity.

Stable nitrogen isotopes. Stable nitrogen isotopes are particularly sensitive to nitrogen cycling within the lake, as well as changes in the relative amount of terrestrial DIN in the lake's nitrogen pool (Peterson and Fry, 1987; Estep and Vigg, 1985; Gu et al., 1994; Gu, 1993c; France, 1995b; See Chapter 2 and Appendix B for a more detailed discussion).

Between 12,000 ^{14}C yr BP and about 9200 ^{14}C yr BP, $\delta^{15}\text{N}$ varies, with lower values from 12,500 to 11,000 ^{14}C yr BP, higher values between 11,000 and 10,200 ^{14}C yr BP, and reduced values 10,200 to 9200 ^{14}C yr BP. C/N ratios as well as C and N concentrations are low throughout this period, indicating low productivity. It is probable that the low $\delta^{15}\text{N}$ values reflect the direct sedimentation of terrestrial N, with little fractionation or recycling. The slight increase in $\delta^{15}\text{N}$ may reflect increased productivity and recycling of N (although the C and N concentrations do not support this),

From 9200 to 8500 ^{14}C yr BP, $\delta^{15}\text{N}$ is elevated, while from 8500 to about 6000 ^{14}C yr BP, $\delta^{15}\text{N}$ decreases. After 6000 ^{14}C yr BP, $\delta^{15}\text{N}$ increases, and remains somewhat stable until the present. C and N concentrations, as well as fluxes are high after about 9500 ^{14}C yr BP, indicating that most of the $\delta^{15}\text{N}$ signal from 9500 ^{14}C yr BP to the present reflects changes of nitrogen cycling. The elevated $\delta^{15}\text{N}$ may be due to increased N recycling prior to sedimentation; while the lowered $\delta^{15}\text{N}$ could reflect higher concentrations of blue-green algae, or increased denitrification with a relative increase of terrestrial N in the nitrogen pool.

From about 8000 ^{14}C yr BP to the present, C/N ratios steadily increase at Birch Lake. Nitrogen isotopes are $>2\text{‰}$ during this period, suggesting little

terrestrial input. *Isoetes* (an emergent aquatic) pollen became common in the Birch G/H pollen record after 8000 ^{14}C yr BP, suggesting a more abundant macrophyte flora in the lake. It is probable, that only after the lake reached its overflow level (see below), and the water level stabilized, could significant macrophyte beds to develop on the shoreline.

Lake level changes. About 13,000 ^{14}C yr BP, silt and sand were deposited in the central cores at Birch Lake, suggesting the lake level was greatly reduced (Figure 4.14). The water level would have been 15-17 m below the modern overflow level (Abbott et al., in prep.). This is also supported by high $\delta^{13}\text{C}$ at this time, indicating reduced ground water flux.

Between about 12,500 ^{14}C yr BP and 12,000 ^{14}C yr BP, shallow water cores show evidence of lacustrine deposition, indicating rapid lake level rise, to as high as 4-6 m below modern levels (Abbott et al., in prep.) (Figure 4.14). In the central core, $\delta^{13}\text{C}$ is reduced at this time, indicating higher ground water flux into the lake.

Lake levels remained at about -5 m for as long as 1000 ^{14}C yr, abruptly decreasing to 9-11 m below modern levels 11,500-11,000 ^{14}C yr BP, and remaining at low levels until a rapid increase about 10,600 ^{14}C yr BP (Abbott et al., in prep.) (Figure 4.14). The exact timing of this lowstand is in some question, however. The lowstand is recognized by shallow water deposits in core D. Correlations of the pollen stratigraphy between cores D and G/H suggest the lowstand occurred simultaneously with the increase in *Artemisia* pollen that is dated between about 11,000 and 10,000 ^{14}C yr BP in core G/H.

Beginning between about 10,600 and 10,000 ^{14}C yr BP, Birch Lake rose rapidly almost to modern levels (-1m), and remained at this high level until about 9800 ^{14}C yr BP (Figure 4.14). An erosional unconformity in the shallowest cores with an upper limiting date of about 8800 ^{14}C yr BP, indicates a brief period of lowered lake level (possibly 2-6 m below modern) between 9800 and 8800 ^{14}C yr BP (Abbott et al., in prep.) (Figure 4.14).

After 8800 ^{14}C yr BP, deepwater lacustrine sedimentation in all Birch Lake cores indicates the lake had reached its overflow stage at this point, and that for the remainder of the Holocene, the lake has not dropped below this level (Abbott et al., in prep.) (Figure 4.14).

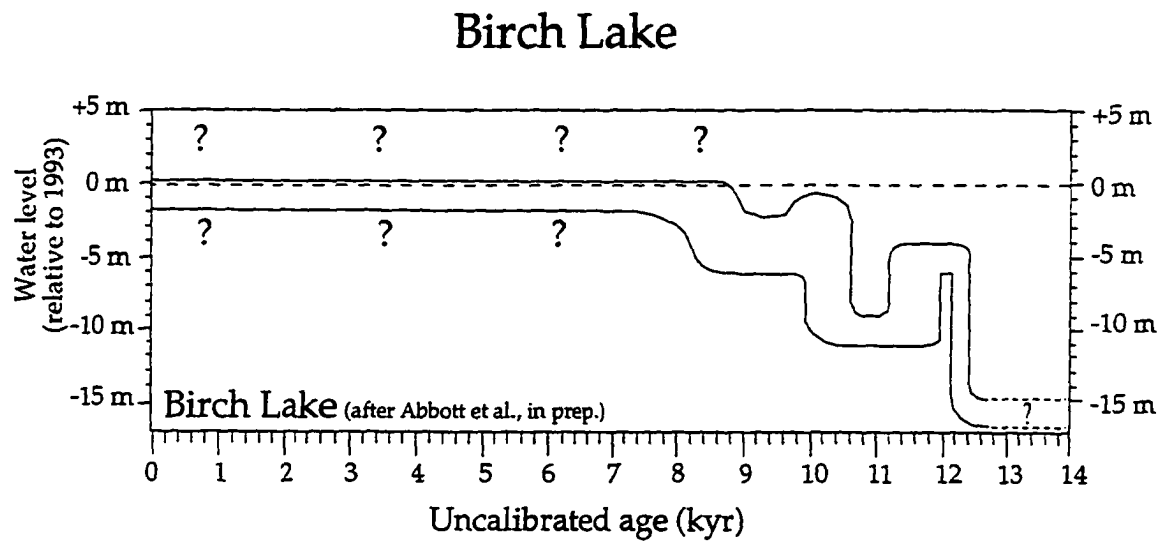


Figure 4.14: Birch Lake lake level changes (after Abbott et al., in prep.).

Summary

Before about 13,000 ^{14}C yr BP, the Birch Lake water level was as much as 15 m lower than modern, ground water inflow was much reduced, and the vegetation was a herbaceous and shrubby tundra.

Beginning about 12,500 ^{14}C yr BP, lake level rose to about 5 m below modern levels, remaining at that level until about 11,600 ^{14}C yr BP. During this period ground water inflow increased, and the vegetation underwent a dramatic shift to a birch-dominated shrub tundra.

The lake level data suggest a period of lower lake levels possibly between 11,000 and 10,000 ^{14}C yr BP. Ground water inflow probably decreased during this period, but may have also remained low until about 9600 ^{14}C yr BP. The vegetation at this time was a shrub tundra, with a short episode of more abundant grasses and *Artemisia* between about 11,000 and 10,000 ^{14}C yr BP.

By about 9800 ^{14}C yr BP, core lithologies suggest the lake level rose to near modern levels. The vegetation was a *Betula* shrub tundra. These data suggest rising effective moisture. Although, carbon isotopes suggest groundwater flux was still at low levels.

The core lithologies suggest a brief decrease in lake level at 9800 to 8800 ^{14}C yr BP indicating a short episode of lowered effective moisture. Common *Populus* on the landscape at the end of the period indicates warm summer temperatures, which is consistent with lowered effective moisture (because of higher evaporation). Carbon isotopes, however indicate rapidly increasing groundwater flux, with peak isotopic values at about 9000 ^{14}C yr BP.

After 8800 ^{14}C yr BP, the Birch Lake water level reached its overflow stage, and the water level has not dropped below the lake outlet since that time. By about 8400 ^{14}C yr BP, *Picea* was present locally, while *Alnus* did not appear until about 6600 ^{14}C yr BP. While the dominant boreal forest species grew at Birch Lake by the middle of the Holocene, variations in their abundances suggest the boreal forest we see today did not develop until the late Holocene.

CHAPTER 5--YOUNGER DRYAS IN ALASKA

Introduction

Roughly between 11,000 and 10,000 ^{14}C yr BP, the pollen records from Windmill and Birch lakes show evidence of climate change in central Alaska. At Windmill Lake, decreases in *Betula* frequency are matched by increases in Poaceae and *Artemisia* pollen. Birch lake shows a similar pattern, although it is less pronounced. These data suggest that, in places, a herb tundra with abundant grasses and *Artemisia* became a more important part of the vegetation.

The period between 11,000 and 10,000 ^{14}C yr BP coincides with the Younger Dryas chronozone, an episode of sharply deteriorating climate that has been documented on the continents surrounding the North Atlantic and elsewhere. This chapter reviews current ideas on the distribution of the Younger Dryas in North America, describes the evidence for climate change at Birch and Windmill lakes, and presents hypotheses explaining the apparent world-wide distribution of this event.

Younger Dryas

The Younger Dryas (YD), originally discovered in northwestern Europe, is dated between about 10,000 and 11,000 ^{14}C yr BP (Mangerud et al., 1974). There has been some discussion on what constitutes the YD: is it only a period of time (YD chronozone) (Mangerud et al., 1974), a biological signal indicating cooling (YD biozone), or some combination of the two? For the purposes of this paper, I consider the YD a period of climatic deterioration between about 10,000 and 11,000 ^{14}C yr BP.

Various sorts of proxy-climatic records from northwestern Europe document the YD. These include pollen (Watts, 1980), beetle fauna (Atkinson et al., 1987), and glacial deposits (Andersen, 1981). It was originally thought that evidence for Younger Dryas cooling decreased in the south of Europe (Watts, 1980), but more recent research indicates significant cooling in southwestern Europe, as well as in the Alps (Amman and Lotter, 1989; de Beaulieu et al., 1994; Lowe et al., 1994).

Marine cores from the North Atlantic preserve strong evidence of Younger Dryas cooling, by increases in $\delta^{18}\text{O}$ and cold-tolerant foraminifera, as well as the presence of ice-rafted debris recording glacier growth in Greenland, (Ruddiman and McIntyre, 1981; Ruddiman, 1987). From Greenland itself, ice cores document abrupt Younger Dryas anomalies in the $\delta^{18}\text{O}$ record, the aerosol dust record, and by changes in ice accumulation rates (Dansgaard, 1989; Taylor, et al., 1993; Alley et al., 1993).

Terrestrial records from eastern North America provide clear evidence of a Younger Dryas cooling. The amount of climatic change is similar to European records, although it was unrecognized until the advent of very fine sampling techniques and highly accurate chronological control. A combination of proxy-climatic datasets (i.e., pollen, sedimentology, Chironomid head capsules, and glacial deposits) provide the strongest evidence of cooling (Mayle et al., 1993a, 1993b; Mayle and Cwynar, 1995; Peteet et al., 1990; Walker et al., 1991; Mott et al. 1986; Stea and Mott, 1989).

In northwest coastal North America, glacial and palynological evidence also indicate Younger Dryas cooling (Reasoner et al., 1994; Mathewes, 1993; Mathewes et al., 1993), but poor radiocarbon control in some of the sites, (especially in Mathewes, 1993), limits their interpretability.

Some marine cores from the temperate and northern Pacific Ocean provide isotopic and foraminiferal evidence of Younger Dryas cooling (Kallel et al. 1988; Kudrass et al. 1991), while others do not (Kiegwin and Gorbarenko, 1992; Morely et al., 1991). In cores where the YD is recognized, the isotopic signal is less pronounced than in the North Atlantic (Zahn, et al., 1991)

Younger Dryas in Alaska. The YD has been documented in only a few localities in Alaska. There is no firm evidence of glacial advance in either the Alaska Range, or the Brooks Range, although in both mountain systems there is evidence for late Wisconsin/early Holocene glacial advances. These advances, however, are not always accurately dated and none can be convincingly dated between 11,000 and 10,000 ^{14}C yr BP (Hamilton, 1986b).

Loess deposits in valleys that drain the northern slope of the Alaska Range may indicate increased aridity and/or increased windiness at about this time (Bigelow et al., 1990). A thin sandy layer within the loess is also preserved at archaeological sites in the Nenana and Teklanika valleys. Radiocarbon ages

underlying the sand average about 11,300 ^{14}C yr BP. Only one site (Dry Creek [Figure 2.1]) has closely overlying ages, but they are somewhat variable (Figure 5.1). A new AMS analysis of a cervid antler from above the sand (10,395 \pm 90 [AA-11729]) is 300 years younger than charcoal analyses from the same layer (Bigelow and Powers, 1994), indicating that the sandy layer had been deposited before 10,400 to 10,700 ^{14}C yr BP.

Pollen records from coastal regions of Alaska indicate climatic change between roughly between 10,000 and 11,000 ^{14}C yr BP. In Southeast Alaska (Pleasant Island [Engstrom et al., 1990; Hansen and Engstrom, 1996]) (Figure 1.2), the YD is marked by a significant reduction in *Pinus contorta* and increases in grasses, sedges, *Artemisia*, and other herbs. Organic content of the sediments also decreases at this time. The radiocarbon chronology is not perfect, but bracketing dates suggest these changes occurred between about 10,800 and 9700 ^{14}C yr BP. The pollen record indicates that a lodgepole pine (*P. contorta*) parkland was replaced by a tundra vegetation with a mix of xeric and mesic species, indicating a significant temperature reduction.

The YD has also been recognized in a paleoecological record from southwestern Kodiak island (Phalarope [Peteet and Mann, 1994]) (Figure 1.2). The YD is marked by an abrupt reduction of ferns (Polypodiaceae) with increases in xerophytic herbs and *Empetrum* seeds. Radiocarbon control is good because several tephra layers assist in age correlations, yielding an age of 10,800 ^{14}C yr BP for the onset of the YD and about 10,000 ^{14}C yr BP for its termination. The abrupt vegetation changes all indicate cooler and drier conditions during the YD.

In southwestern Alaska, two records may show evidence of a YD climatic reversal between about 10,800 and 9800 ^{14}C yr BP (Grandfather and Ongivinuk [Hu et al., 1995]) (Figure 1.2). The YD subzone is marked by a brief, but abrupt return to herb tundra. Prior to the YD, *Betula* accounts for 20% of the pollen, at the onset of the YD, *Betula* drops to almost nil, and rapidly recovers by about 9500 ^{14}C yr BP. At the same time, the abundance of herbaceous species increased or remain the same. Indicators of lake productivity (organic content, biogenic silica, and *Pediastrum* nets) also declined at this time, suggesting a catchment-wide response to YD cooling.

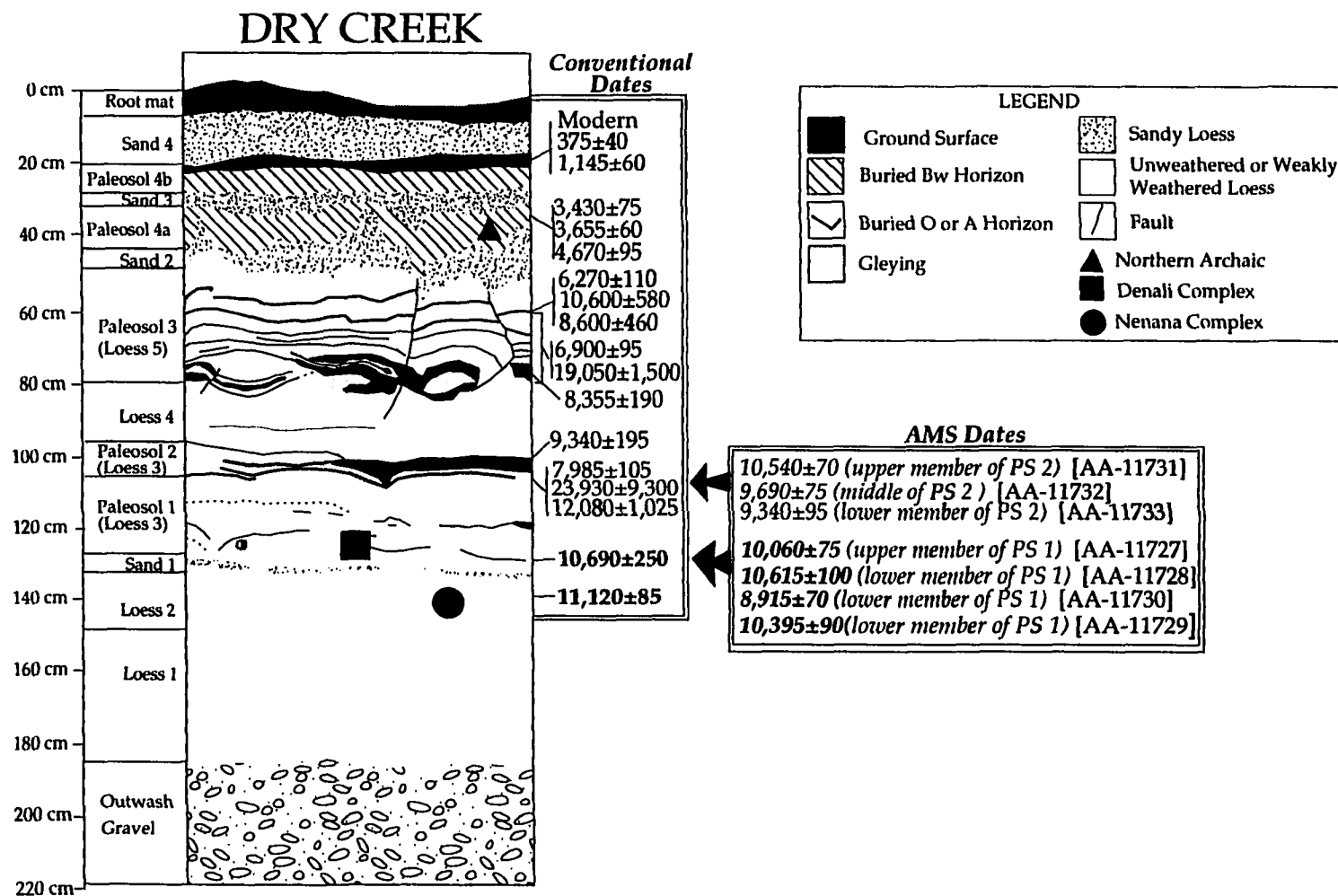


Figure 5.1: Schematic drawing of the Dry Creek loess profile with associated conventional and AMS determinations. Dates in bold type relate to the Younger Dryas event.

New Evidence from Central Alaska

Isotopes and lake level changes. Carbon isotopes at Birch lake are at a relatively high level between 9800 and 11,200 ^{14}C yr BP, while at Windmill, two samples at about 10,400 ^{14}C yr BP show higher $\delta^{13}\text{C}$ (Figure 5.2). These changes in $\delta^{13}\text{C}$ either reflects a change in the water chemistry entering the lakes, or a change in the carbon cycle within the lakes. Carbon and nitrogen concentrations at both lakes are low during this time, suggesting productivity changes aren't driving the isotopic signal. Instead, the high $\delta^{13}\text{C}$ signal reflects changes the lakes' DIC as a result of changes in the water budget. A reduction in ground water flux relative to precipitation probably explains the increased $\delta^{13}\text{C}$ of the sediments.

At Birch Lake, a lowered groundwater flux is supported by stratigraphic changes in the shallow cores that indicate the water level dropped five to seven meters about 11,500 to 10,600 ^{14}C yr BP (Abbott et al., in prep.) (Figure 4.14). At Windmill Lake there is no specific evidence of lowering lake level at this time, although the water level was still about 4 m below modern (Figure 2.16).

Pollen. The pollen data indicate short-term changes in the landscape and vegetation, suggesting increasing aridity and/or cooling. At both Windmill and Birch lakes, between about 10,800 and 10,100 ^{14}C yr BP, *Betula* decreases, while *Artemisia* and *Poaceae* increase (Figures 5.3 and 5.4). To address whether these changes are significant, the datasets were ordinated together, using the samples from the base of the cores up to about 8400 ^{14}C yr BP, when *Picea* becomes common on the landscape. This group of samples was chosen because it is important to see how the YD samples differed from both the earlier herb zone samples and from the birch zone samples that surround them. Prior to the analysis, the Birch and Windmill pollen frequencies were calculated omitting *Picea* and *Alnus*. The ordination is a PCA on square root transformed percentage data (Figure 5.5).

When this analysis is examined in detail, it is clear that the YD samples from Windmill Lake revert towards the herb zone (Figure 5.6). Initially, most of the change is on the second axis, although during the YD, the first axis also becomes important. At the end of the YD, the samples move towards their pre-YD position on both axes.

$\delta^{13}\text{C}$ at Windmill and Birch Lakes

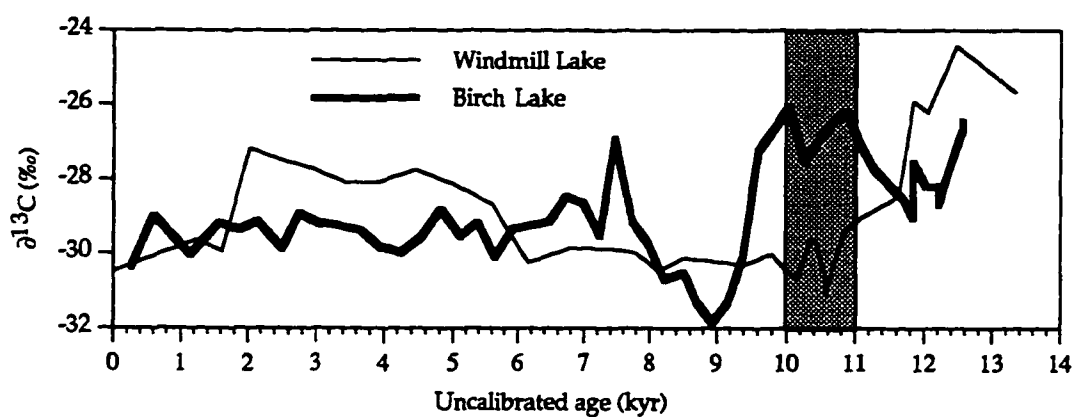


Figure 5.2: $\delta^{13}\text{C}$ at Windmill and Birch lakes. Shading marks the Younger Dryas chronozone.

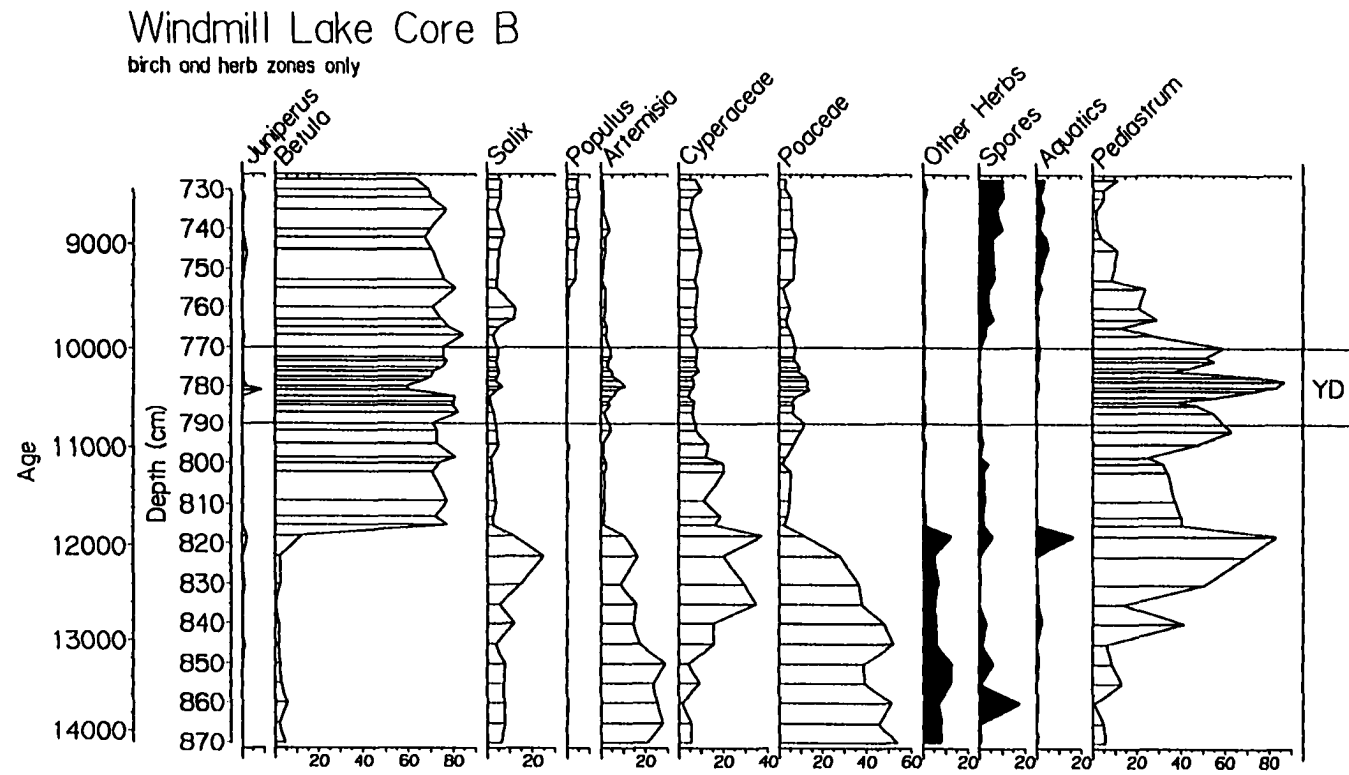


Figure 5.3: Windmill Lake birch and herb zones, with Younger Dryas marked.

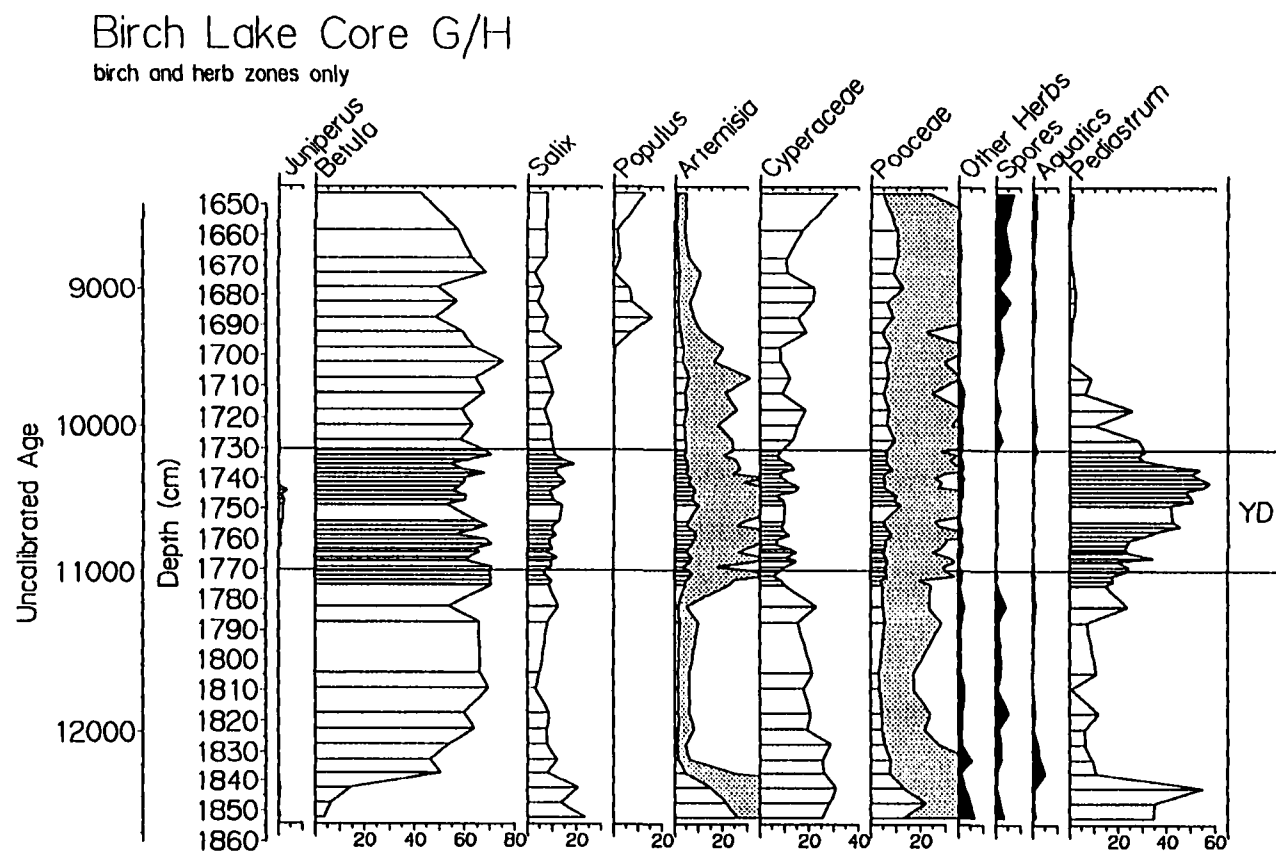


Figure 5.4: Birch Lake birch and herb zones, with Younger Dryas marked. Shading indicates 5x exaggeration.

Windmill and Birch Lakes

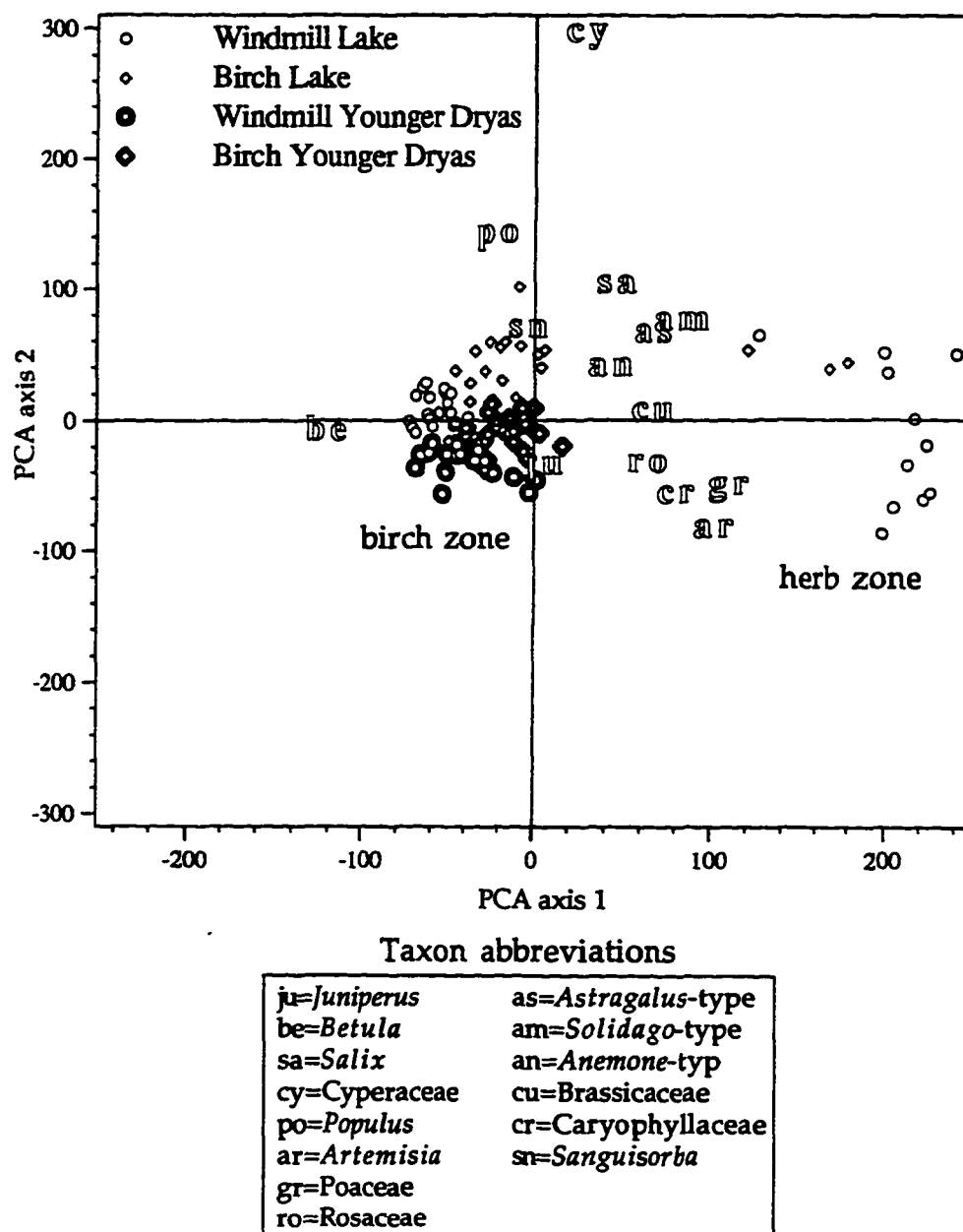


Figure 5.5: PCA biplot of birch and herb zones from Windmill and Birch lakes. Younger Dryas samples are marked.

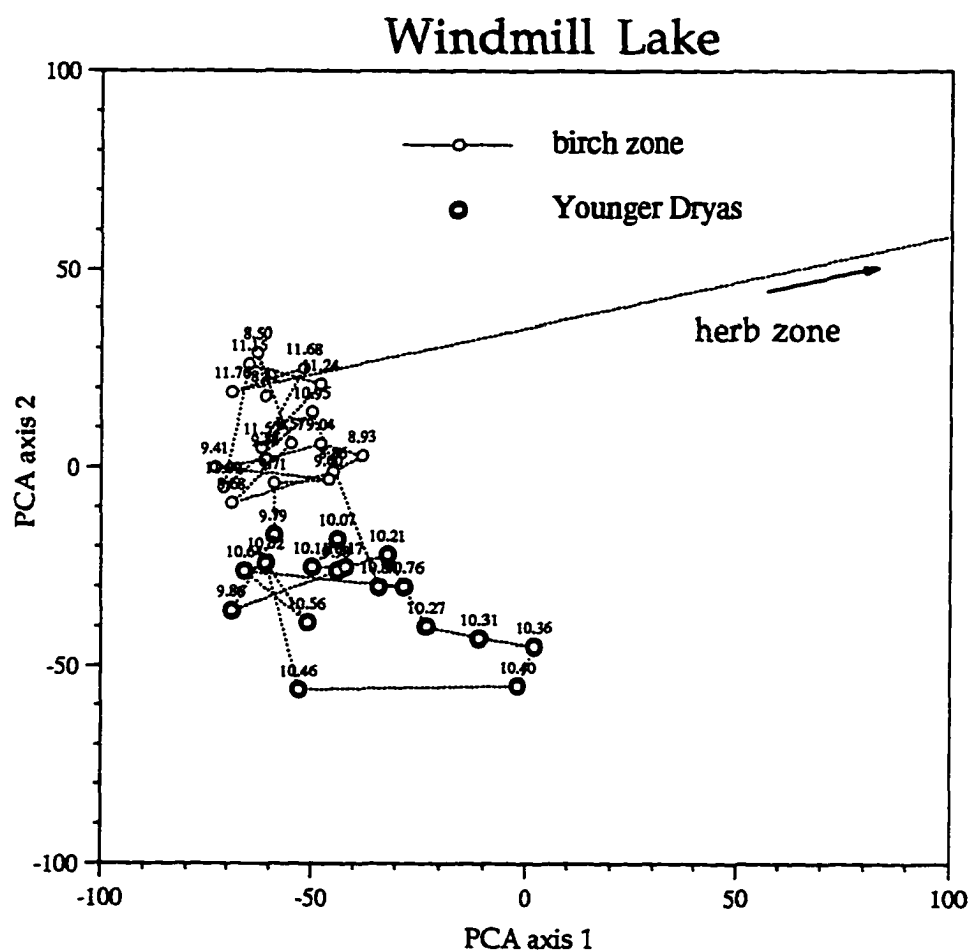


Figure 5.6: PCA biplot of Windmill Lake birch zone with Younger Dryas samples marked. Numbers indicate uncalibrated age of each sample.

The YD samples at Birch Lake also revert towards the herb zone, but the pattern is not as clear as at Windmill Lake (Figure 5.7). At Birch Lake, most of the change is on the second axis. The separation of the YD samples from the surrounding samples is not as great as at Windmill Lake. This suggests that the changes in the pollen spectra are indeed smaller at Birch Lake and this difference probably reflects a more subdued vegetation response to regional climate changes.

T-tests ($P \leq 0.05$, two-tailed, assuming unequal variances) comparing mean PCA scores of the YD sample group in each lake with those coming immediately before and after the Younger Dryas suggest both that the YD group is significantly different from the preceding and succeeding samples, and that the samples before and after the YD are not significantly different from each other (Table 5.1). The T-tests also indicate that at Windmill Lake, axes 1 and 2 delineate the YD samples, while at Birch Lake, only axis 2 does.

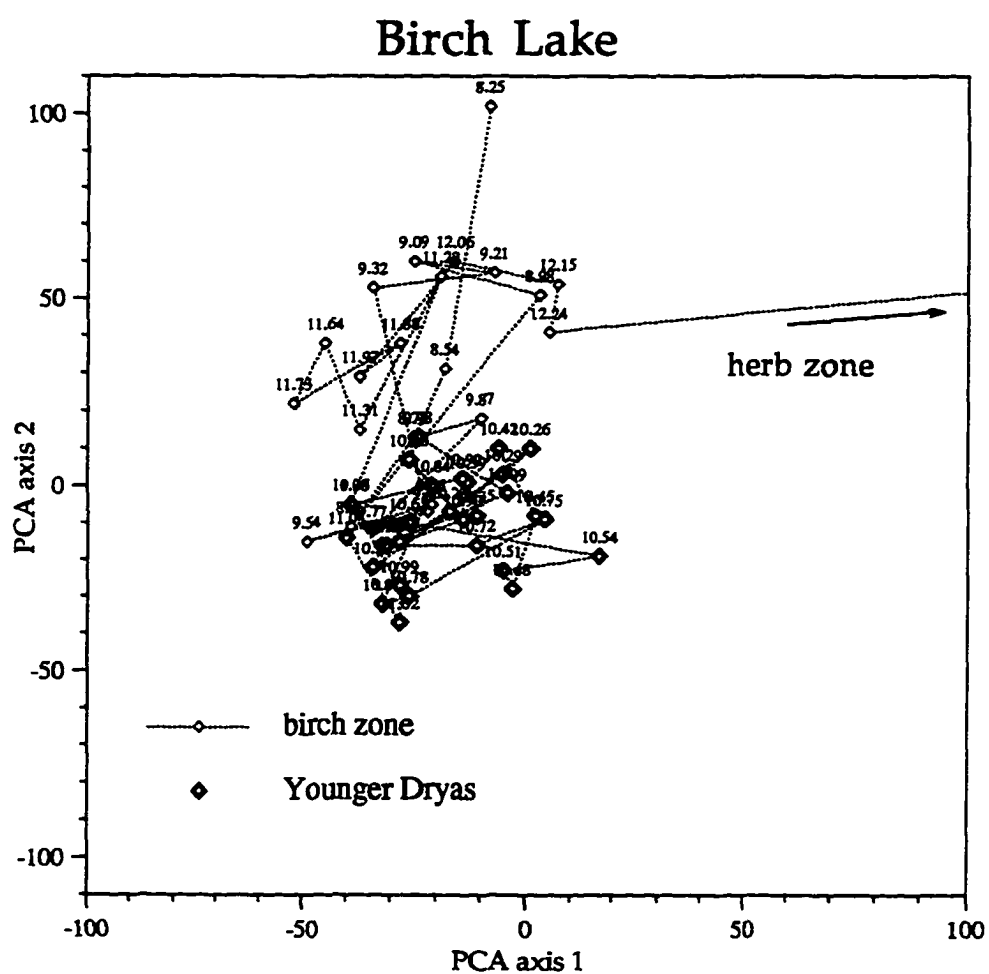


Figure 5.7: PCA biplot of Birch Lake birch zone with Younger Dryas samples marked. Numbers indicate uncalibrated age of each sample.

Table 5.1: T-test of YD samples with BYD (before YD) and AYD (after YD) samples. Bold type indicates the groups are significantly different ($P \leq 0.05$).

	Windmill	Birch
Variables	axis 1: YD;BYD	axis 1: YD;BYD
m,n;df¹	17,7;22	29,10;12
t	-3.227	-1.313
t_{critical}	-2.080	-2.179
P	0.004	0.214
Variables	axis 1: YD;AYD	axis 1: YD;AYD
m,n;df	17,11;25	29,13;23
t	2.764	-1.076
t_{critical}	2.064	-2.074
P	0.01	0.294
Variables	axis 1: BYD;AYD	axis 1: BYD;AYD
m,n;df	7,11;14	10,13;16
t	-0.700	-0.514
t_{critical}	-2.145	2.132
P	0.496	0.615
Variables	axis 2: YD;BYD	axis 2: YD;BYD
m,n;df	17,7;12	29,10;12
t	9.227	6.611
t_{critical}	2.201	2.201
P	0.000	0.000
Variables	axis 2: YD;AYD	axis 2: YD;AYD
m,n;df	17,11;23	29,13;13
t	-8.523	-3.481
t_{critical}	-2.069	-2.160
P	0.00	0.004
Variables	axis 2: BYD;AYD	axis 2: BYD;AYD
m,n;df	7,11;12	10,13;19
t	1.994	0.753
t_{critical}	2.179	2.093
P	0.069	0.461

Comparisons with modern pollen. Today, in central Alaska, *Artemisia* is most common on south-facing bluffs. Modern pollen from Bonanza Bluff, located near Fairbanks (Figure 2.1), was collected over the course of one summer (see Appendix C). The pollen rain at this site is unusual, because of

$$^1df = (S_1^2/m + S_2^2/n) / ((S_1^2/m) / (m-1) + ((S_2^2/n) / (n-1))$$

where

S_1 = variance of group 1, m = # of observations in group 1

S_2 = variance of group 2, n = # of observations in group 2

the high frequencies (mean=19%; range=1%-70%) of *Artemisia* pollen. Today, in northern North America, only the fell fields of the northern Yukon (Cwynar, 1982) and parklands/grasslands of southern Canada (Ritchie, 1987) have frequencies this high.

To compare the Bonanza Bluff modern pollen spectra with the Birch and Windmill fossil spectra, all three datasets were ordinated together using PCA. The pollen frequencies from the three sites were calculated omitting *Picea* and *Alnus*, as well as other minor boreal forest taxa (such as *Larix*, *Myrica*, *Shepherdia*, *Viburnum*, *Linnea*, and *Cornus*). As with earlier ordinations, the analysis is based on terrestrial taxa with a maximum abundance $\geq 2\%$, and in the case of the lake cores, which also occurred >5 times. Table 6.2 lists the taxa used in this analysis

Table 5.2: Taxa included in Bonanza, Windmill, and Birch ordination.

<i>Juniperus</i>
<i>Betula</i>
<i>Salix</i>
<i>Populus</i>
Ericales
<i>Artemisia</i>
Cyperaceae
Poaceae
<i>Aconitum</i>
<i>Pulsatilla</i> + <i>Anemone</i>
Rosaceae undiff.
Caryophyllaceae
<i>Ambrosia</i> -type
<i>Astragalus</i> -type
Brassicaceae
<i>Sanguisorba</i>

The biplot of the sample scores indicates that the ordination separates the herb zone from the birch zone in the fossil samples, while the modern samples lie between them (Figures 5.8 and 5.9). Compared with the fossil samples, the Bonanza Bluff samples are more scattered; this is because pollen collected in pollen traps is inherently more variable than pollen preserved in lakes (Faegri and Iversen, 1989).

Windmill Lake and Bonanza Bluff

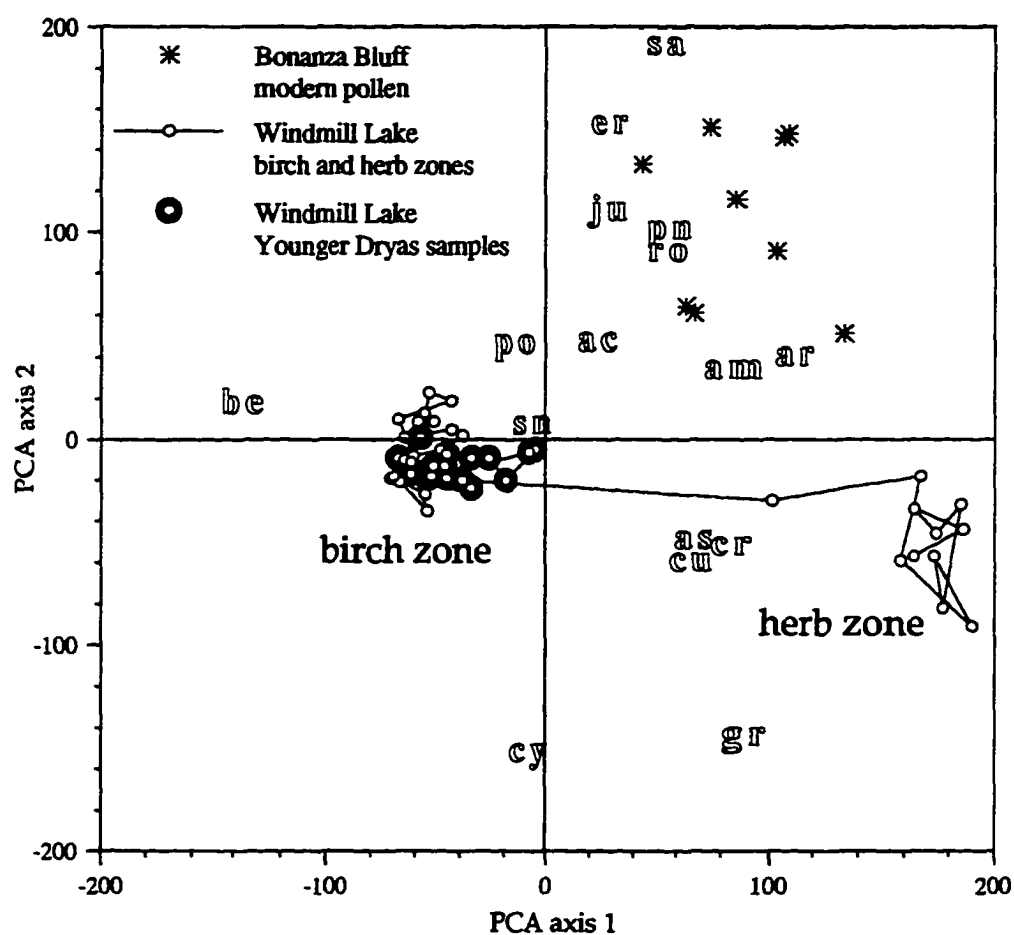


Figure 5.8: Windmill Lake and Bonanza Bluff PCA ordination.

Birch Lake and Bonanza Bluff

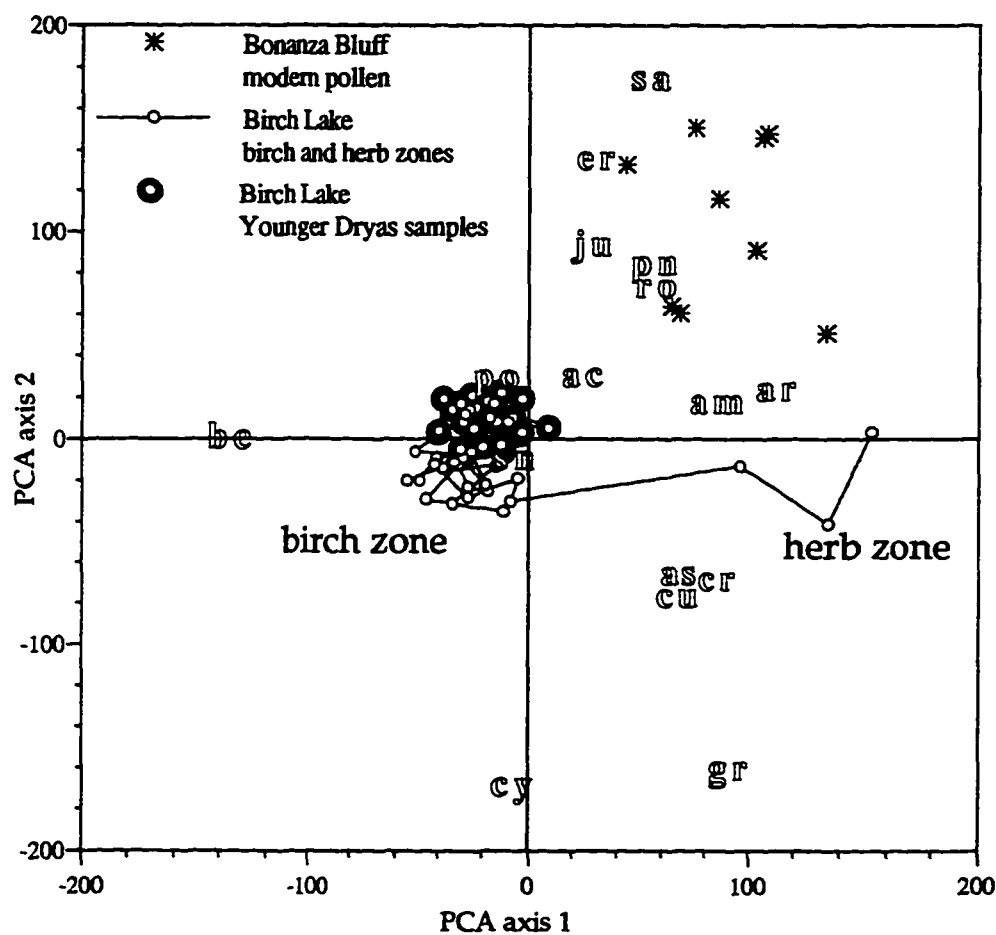


Figure 5.9: Birch Lake and Bonanza Bluff PCA ordination.

When this analysis is examined in detail, it appears that the YD samples from Windmill Lake are more similar to the herb zone samples than to the Bonanza Bluff samples, while at Birch Lake, the reverse is true.

Discussion

The Younger Dryas at Birch and Windmill lakes Pollen data from Birch and Windmill lakes suggest an expansion of xeric and cold-adapted taxa in both the Tanana floodplain as well as in the Alaska Range. *Betula* was undoubtedly still common on the landscape throughout central Alaska, but the *Artemisia* and grasses may have expanded into areas sensitive to increasing aridity and/or cooler temperatures. At low elevations, the *Artemisia* and herbs may have been more common on south-facing bluffs and well-drained floodplains. At higher elevations, the *Artemisia* and herbs in the herb tundra may have expanded into the *Betula* shrub tundra.

At Birch Lake, the *Artemisia* peak is probably coincident with the lake low stand and elevated carbon isotopes (Figure 5.10), suggesting that the aridity was severe enough to lower the lake level as well as favor xerophytic taxa over mesophytic taxa. Comparison of the Birch Lake pollen record with the modern pollen at Bonanza Bluff suggests that the YD samples are more similar to Bonanza Bluff (a very arid locality today) than they are to the previous herb zone.

Birch Lake, because it is situated away from the mountains and on the Tanana floodplain, may have been more severely affected by aridity changes than Windmill Lake. Today, moisture entering central Alaska originates either in the Gulf of Alaska or Bering Sea, although most of the moisture from the Gulf of Alaska is blocked by the Alaska Range (Bowling, 1979). During the YD, significantly less moisture was probably being advected into central Alaska, because of lowered sea surface temperatures, and because the Bering Land Bridge was still partially exposed, and less moisture could come from that source.

The *Artemisia* rise at Windmill Lake is more abrupt and slightly later than at Birch Lake (Figure 5.10); the isotope signal, while slightly elevated, is not of the same magnitude as Birch Lake. Comparison of the Windmill Lake YD samples with the modern pollen from Bonanza Bluff suggests the YD

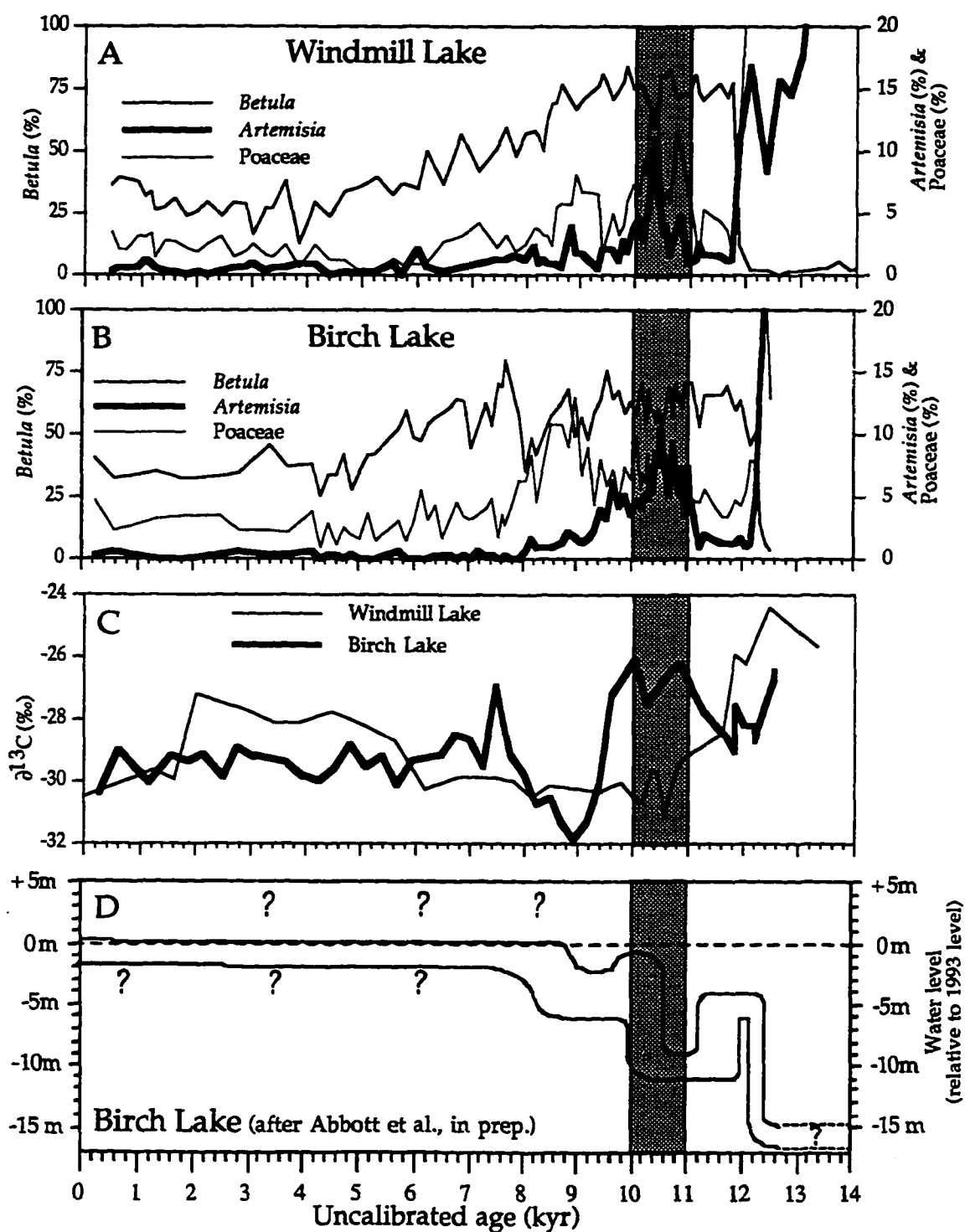


Figure 5.10: (A) Windmill Lake pollen, (B) Birch Lake pollen; (C) Birch Lake $\delta^{13}\text{C}$; (D) lake level changes at Birch Lake. Shading marks the Younger Dryas chronozone.

samples are more similar to the cold and arid herb zone than to arid Bonanza Bluff. These data suggest that aridity may not have been the main climatic variable affecting Windmill Lake and its watershed during the YD chronozone.

Windmill Lake, because of its location in the Alaska Range, receives moisture from the Gulf of Alaska today. It is possible, that during the YD, the changes in moisture had a smaller effect on the vegetation at this site, because it still received moisture from the North Pacific (albeit significantly less than today). However, because of its elevation (640 m), this site was probably quite sensitive to changes in temperature. Thus, the vegetation changes associated with the YD may reflect cooler temperatures as well as slightly drier conditions.

Younger Dryas climate. Proxy climatic records from interior Alaska unequivocally indicate very dry climates during the YD chronozone. YD-aged loess deposits at archaeological sites in the Nenana and Tanana valleys are unweathered, (Powers and Hoffecker, 1989; Bigelow, 1991; Yesner et al., 1992) suggesting ongoing loess deposition and poor soil formation. A sandy layer found at some sites also indicates abundant source material (possibly due to sparse vegetation cover), and increased windiness during the YD (Bigelow et al., 1990).

In the southern Yukon, a pollen record (Kettlehole Pond) (Figure 1.2) from the rain shadow of the coastal mountains contains very high *Artemisia* frequencies during the YD chronozone (Cwynar, 1988), which also indicate increased aridity. Although the author suggested this was probably due to higher summer temperatures instead of reduced precipitation.

Pollen records from coastal Alaska show cooler temperatures as well as increased aridity. On Kodiak island, the loss of the *Polypodiaceae* indicates both drying and cooling, while the increase in *Empetrum* seeds suggests drying (Peteet and Mann, 1994). In southeast Alaska, the loss of trees, shrubs, and ferns (*Dryopteris*) along with the gain of herbaceous taxa, especially *Artemisia*, indicate cooling and drying (Hansen and Engstrom, 1996). In southwestern Alaska, a reduction in *Betula* and *Polypodiaceae* also show cooler and/or drier conditions during the YD (Hu et al., 1995).

Based on an admittedly small dataset, it appears that coastal sites, as well as those in the interior that receive coastal moisture (i.e., Windmill Lake), experienced extensive cooling, as well as increased aridity, while the interior sites (Birch Lake, Kettlehole Pond) were affected mainly by increasing aridity. *Younger Dryas and the Pacific rim.* On a global scale, the intensity of the climatic change appears strongest in regions surrounding the North Atlantic, indicating this region is the source area for the YD. If this is the case, and if there was a YD event in Alaska, then we must explain how it was propagated from the Atlantic to the Pacific.

Researchers have proposed that discontinued and/or altered formation of North Atlantic deep water (NADW) triggered YD cooling in regions surrounding the North Atlantic (Broecker et al., 1989; 1990; Broecker, 1995). NADW is part of a global "conveyor belt" of oceanic circulation. In the Atlantic Ocean, warm subtropical waters travel northwards to about 60° N, as these surficial waters cool, they sink, releasing latent heat into the atmosphere. It is thought that this circulation supplements the yearly insolation in the North Atlantic region by about 25% (Broecker, et al, 1990). Thus, an interruption in deep water formation would have a significant effect on the climate in the regions surrounding the North Atlantic.

Density changes of the surface waters can halt the conveyor belt, because if the warm surface waters remain less dense (i.e., less salty) than the underlying water, they no longer sink, thus starving the system. During the Younger Dryas, a less dense, freshwater cap on the North Atlantic from the melting Laurentide ice sheet, may have stopped the deepwater cycle (Broecker et al., 1989). However sea level changes indicate a reduction in glacial meltwater during the YD (Fairbanks, 1989; Edwards et al., 1993), suggesting that if deepwater formation did cease during the YD, then other factors related to the inherent instability of the North Atlantic circulation were the cause (Broecker, 1995).

Halting the global conveyor belt alone probably does not explain the evidence for YD cooling on the Pacific rim, because little deep water forms there today. In addition, the Pacific Ocean is inherently more stable than the Atlantic Ocean (Zahn et al., 1991). How the changes in the North Atlantic region connect with climate changes in the North Pacific region is poorly

understood. The absence of a significant time-lag between these two regions indicates that any connection through the marine deep water system is unlikely because it takes about 1000 years for deep water to flow from the North Atlantic to the Pacific (Kennett and Ingram, 1995). This indicates that global teleconnections of YD cooling must have been mediated through the atmosphere and/or through changes in marine surface water temperatures.

The atmospheric linkage may have been through two different mechanisms. First, decreases in greenhouse gases such as methane and CO₂ during the YD possibly lowered atmospheric temperatures. Ice cores from Greenland suggest both gasses are at low levels both during the full glacial and during the YD, although they alone were not enough to cause the 2°C to 7°C shift in atmospheric temperatures that have been documented for the Younger Dryas (Chappellaz, et al., 1993; Paterson and Hammer, 1987).

Second is an increase in volcanism just prior to the onset of the YD. Five major tephra deposits around the world have been dated to about 11,100 to 11,300 ¹⁴C yr BP. It is possible these eruptions, by spewing particulates into the atmosphere, initiated a short sharp period of climatic cooling. However, increased volcanism alone would only cool the globe for a short time, so an additional forcing is needed to explain the 1000 yr-long YD event.

The marine linkage may have been through changes in sea surface temperatures. The upper mixed layer of the ocean can travel rapidly (c.f. El Nino events), indicating that a cooler North Atlantic sea surface may have quickly propagated to the Pacific ocean (especially if mediated through the atmosphere). The duration and intensity of the climate change in Alaska suggest that the North Pacific Ocean must have been involved in the YD event. A cooler North Pacific would have cooled the coastal areas, as well as reducing evaporation and the transport of moisture into the interior. Sea surface temperatures need not have been significantly cooler to cause the documented climatic changes. A preliminary climate model simulation suggests that a 2°C shift in North Pacific sea surface temperatures would be sufficient to cool the northern hemisphere and increase snow fall in coastal areas (Peteet and Mann, 1994).

Summary

Lacustrine and vegetation evidence from Birch and Windmill Lakes indicate a significant change in the regional water table and plant distributions during the YD. All the evidence suggests increasing aridity and/or cooling roughly between 11,000 and 10,000 ^{14}C yr BP. Interior Alaska may be sensitive to changes in effective moisture because this region has low annual precipitation, and even a small reduction in effective moisture would significantly affect the vegetation.

Differences in the proxy records from the Alaska Range and the Tanana floodplain suggest the YD affected these regions differently. In the Alaska Range, the vegetation changes markedly, possibly reflecting an increase in alpine, as well as xerophytic vegetation. In the Tanana floodplain, the vegetation change is not as dramatic, probably reflecting limited increases in low elevation xerophytic taxa.

The absence of a significant lag in the onset of the YD in the North Pacific region suggests that the climate change must have been mediated by changes in the atmosphere or surface ocean waters. The duration of the YD (almost 1000 ^{14}C yr) requires that the Pacific Ocean be cooled during the YD, because atmospheric changes are comparatively transient, or not strong enough to generate the amount of climatic cooling indicated by the various proxy records.

CHAPTER 6—SUMMARY AND SPECULATIONS

Introduction

This chapter summarizes the vegetation and landscape changes documented in the records from Windmill, Dune, and Birch Lakes. Afterwards follows a discussion of how sea level changes may have affected the climate of central Alaska. Finally, I present speculations of how people could have been affected by climate and landscape evolution since deglaciation.

Vegetation and Landscape Evolution in Central Alaska

Before about 12,000 ^{14}C yr BP, low pollen influx in the earliest levels of the Windmill and Birch pollen records indicate a poorly vegetated landscape. Sparse clumps of *Artemisia*, Cyperaceae, and Poaceae grew in the exposed sites, while *Salix* was probably concentrated in alluvial settings or in areas with a thick snow cover. The vegetation surrounding the lakes was probably a mixture of Cyperaceae, *Salix*, and other mesophytic species. Many of the taxa present before 12,000 ^{14}C yr BP were primary successional species that grow well on disturbed ground, indicating abundant exposed mineral soil. Low lake levels (Windmill and Birch) or completely dry lakes (Dune) indicate very low effective moisture at this time. This is consistent with the climate model simulations suggesting that ice sheet circulation effects during the full-glacial would have diverted the jet stream southward resulting significantly reduced moisture advection into central Alaska (Bartlein et al., 1991; Kutzbach et al., 1993)

About 11,800 ^{14}C yr BP, the vegetation in central Alaska changed significantly with the addition of *Betula* (although the herbaceous species from the previous period continued, often more abundantly). The *Betula* pollen probably represents shrub or dwarf varieties that grew in protected sites, or localities with a thick snow cover, while the herbs/forbs grew on well-drained, exposed sites. The addition of *Betula* to the regional vegetation suggests greater warmth and effective moisture, which is also supported by higher summer insolation (Berger, 1978) and rising lake levels. Climate

model simulations indicate that by about this time, ice sheet circulation effects had relaxed somewhat, and the jet stream had migrated northwards close to its modern position, resulting in higher moisture flux into central Alaska (Bartlein et al., 1991; Kutzbach et al., 1993).

Between about 11,000 and 10,000 ^{14}C yr BP, transient changes in vegetation reflect a significant climatic change related to the Younger Dryas event. Increases in herbaceous taxa indicate a shift to drier and cooler conditions. The lake level data from Birch Lake, suggest a 10 m depression in water level possibly from about 11,000 to 10,000 ^{14}C yr BP, indicating reduced effective moisture.

From about 10,000 ^{14}C yr BP to about 8500 ^{14}C yr BP, *Populus* became more common around Windmill and Birch lakes. The arrival of *Populus* was slightly earlier and more abundant at Dune lake, probably due to warm and dry local conditions. The increased *Populus* probably reflects increased early summer insolation. Effective moisture may have decreased as indicated by a lowered lake level at Birch Lake between 9800 and 8800 ^{14}C yr BP. The vegetation at this time was *Betula*-dominated, which probably occupied the cooler mesic sites, while *Populus* grew mainly in riparian settings with good drainage.

By about 8500 ^{14}C yr BP, *Picea* was present, but probably fairly sparse in the region. It didn't grow locally around the lakes until about 8200 to 8000 ^{14}C yr BP. The lake level data, especially from Dune Lake, indicates a significant increase in effective moisture with the appearance of spruce in the region. Summer insolation was also at maximum values at this time, which probably also aided the rapid *Picea* dispersion. The higher effective moisture at this time may have been a critical factor allowing spruce's rapid colonization of the landscape. The climate model simulations do not adequately explain why effective moisture increased, suggesting that regional (as opposed to hemispheric) changes in atmospheric circulation caused the increase in effective moisture (this is discussed more in the following section).

By about 7000 to 6500 ^{14}C yr BP, *Alnus* probably grew locally around Windmill, Birch, and Dune lakes. With the appearance of *Alnus*, the final species of the modern boreal forest was present in central Alaska. The forest,

however, was probably somewhat different from what we see today. *P. glauca* was probably more common than today, and wide spread paludification had not yet begun. Higher than modern lake levels at this time suggest continued increases in effective moisture.

During the mid-Holocene, frequencies of the major pollen taxa were highly variable at Windmill and Dune lakes and probably reflect local perturbations in the vegetation. At Windmill Lake, a combination of repeated forest fires and insect attacks may have episodically decimated the spruce forest. At Dune Lake, local fluctuations in effective moisture may have affected the local distribution of birch, spruce, and alder.

At all three lakes *Picea* pollen frequencies and influx were higher than modern during the mid- to late Holocene, suggesting that spruce was more common than it is today, and subsequently declined to modern levels. This change may mark the true development of the modern boreal forest. Climate was apparently cooling at the end of the Holocene, as reflected in glacier growth and the southward shift of treeline (Hamilton, 1986b; Calkin, 1988; Ritchie et al., 1983; Moser and MacDonald, 1990). This probably resulted in increased paludification and the widespread development of ice-rich permafrost in central Alaska. As muskegs expanded across the landscape, spruce that were formerly growing in the low-lying areas would have thinned or locally disappeared, although the spruce on hillsides could continue as before.

Effects of Sea Level Change on Effective Moisture in Central Alaska

Due to changes in atmospheric circulation, shoreline changes associated with rising and falling sea levels could have had a major effect on the climate of central Alaska.

During the full-glacial, when sea level was 100 to 120 m below modern (Fairbanks, 1989; Edwards et al., 1993), the Bering Land Bridge (BLB) was subaerially exposed. At a 120 m sea level depression, the southern coast of the BLB would have extended from the northern Alaska Peninsula, trending to the northwest, and encountering the Russian coast south of the Gulf of Anadyr (Figure 1.1). In the Arctic Ocean, the landmass would have extended

several 10s of km north of the current Alaskan and Siberian shorelines. The presence of such a large landmass would have had a significant effect on regional climate.

Today, Alaska's climate is driven by the opposing forces of pressure ridges and troughs over Alaska, Siberia, and the North Pacific Ocean. Central Alaska receives most of its annual precipitation during the summer and early autumn. A strong pressure ridge located over eastern Alaska and western Canada flattens over the course of the summer, surface air circulation then shifts to a southwesterly direction, which advects significant moisture from the southern Bering Sea into central Alaska (Bowling, 1979).

During the full-glacial, the exposed BLB would have reduced the amount of moisture being advected (because less moisture can be evaporated from land than from the sea, and sea surface temperatures were cooler), as well as increasing the distance from central Alaska to the moisture source. This, combined with the circulation changes associated with the large Laurentide ice sheet (Kutzbach et al., 1993) resulted in very little moisture reaching central Alaska during the full-glacial.

As the BLB was gradually flooded over the course of the early Holocene, increasing moisture was advected into central Alaska because moisture sources were closer, sea surface temperatures increased, and the atmospheric circulation shifted to the modern pattern. Changes in the vegetation and lake level reconstructions indicate that the periods centered about 12,000 ^{14}C yr BP (higher lake levels and *Betula* rise) and about 8500 ^{14}C yr BP (higher lake levels and *Picea* rise) reflect significant changes in the moisture regime in central Alaska. To what extent was the flooding of the BLB implicated in these changes?

Bathymetric features of the Bering Sea and Bering Strait control the timing of the flooding of the BLB. Three sills are important for this discussion. The deepest sill is the Bering Strait (Figure 1.1), with a maximum depth of 52 m. Anadyr Strait, separating St. Lawrence Island from the Chukotsk Peninsula, (Figure 1.1) has a maximum depth of 46 m, and Shpanberg Strait (Figure 1.1), separating St. Lawrence Island from the Yukon Delta, has a maximum depth of 32 m (McManus and Creager, 1984).

One must be careful extrapolating modern bathymetry into the past. Sedimentation rates vary across the BLB, from almost 9 m to less than 1 m since 14,000 ^{14}C yr BP (McManus and Creager, 1984), indicating that in some localities (especially near river deltas), flooding would occur sooner than it appears using the modern bathymetry. In addition, due to the sediment load, these same localities may also be down-warped today, which would partially offset the effects of the thicker sediment package.

After maximum sea level depression of 120 m at 18,000 ^{14}C yr, the sea level rose rapidly, with a particularly rapid event between about 12,500 and 11,500 ^{14}C yr BP, when the sea rose about 20 m in less than 1000 ^{14}C yr. The rate of sea level rise subsequently decreased somewhat between 11,500 and 10,000 ^{14}C yr BP. At 10,000 ^{14}C yr BP, the rate increased slightly, but began to taper off by about 7000 ^{14}C yr BP. Since about 6000 ^{14}C yr BP, sea level has only risen about 10 m (Fairbanks, 1989).

Between 12,500 and 11,500 ^{14}C yr BP, despite the rapid sea level rise, the amount of flooding on the Alaskan side of the BLB was comparatively small, although the Gulf of Anadyr on the Russian side became partially flooded. Between 11,500 and 9500 ^{14}C yr BP, the Gulf of Anadyr took on its modern configuration, Bristol Bay (Figure 1.1) was partially flooded, and the Anadyr and Bering straits were narrowly breached.

From 9500 to 8500 ^{14}C yr BP, a large proportion of the remaining BLB was flooded, including most of Bristol Bay and the shelf region from the mouth of Norton Sound to St. Lawrence Island, while the Bering Strait became as wide as it is today. Along the Yukon-Kuskokwim delta, however, a shelf about 5 km wide may have persisted until about 6000 ^{14}C yr BP, when it and Norton Sound were flooded.

Lake level data from Windmill and Birch lakes indicate a significant increase in effective moisture coincident with the *Betula* rise about 11,800 ^{14}C yr BP. Only the western part of the BLB was submerged at this time (despite the rapid sea level rise), suggesting that the increased effective moisture in central Alaska was not due to submergence in the BLB. It may in fact, have more to do with weakening of the Laurentide anticyclone and the northerly

shift of the jet stream, which would have brought North Pacific moisture into central Alaska.

By 8500 ^{14}C yr BP, when *Picea* appeared and effective moisture increased in central Alaska, the Bering Sea had taken on its modern configuration, (with the exception of Norton Sound and the Yukon-Kuskokwim delta, which were still exposed), suggesting that the modern marine and atmospheric circulation developed at this time. Once the modern circulation was in place, then increasing amounts of moisture could be advected from the southern Bering Sea into central Alaska. To what extent the documented increases in effective moisture reflect the final flooding of the BLB and the onset of modern atmospheric circulation are conjectural, but the synchronicity of the events is intriguing.

Climate Change and People in Central Alaska

Introduction. This section examines how the vegetation and climate changes described in earlier chapters may have affected the people living in central Alaska. This inevitably involves some speculation on what it was like to live in Alaska in the past. The low number of archaeological sites, combined with few faunal remains make it impossible to state definitively what happened to the people. The low density of sites makes it difficult to assess settlement patterns or to separate functional and cultural differences between assemblages (c.f. Binford, 1983). The paucity of faunal remains make it difficult to know what the people actually hunted throughout the year (c.f. Binford, 1978). For example, moose are poorly represented in archaeological sites (Yesner, 1989), is this because: i) people were not eating moose, ii) the moose were processed off-site, so few remains were recovered in the excavation, or iii) the moose hunting camps have been overlooked in archaeologists' survey strategies.

The paleoecological reconstructions from Birch, Windmill and Dune lakes suggest four periods of significant climate change since deglaciation. At 12,000 ^{14}C yr BP, climate warmed and effective moisture increased, although conditions were still cooler and drier than today. Between about 11,000 and 10,000 ^{14}C yr BP, climate became increasingly arid and/or cooler. Finally, by

8500 to 8000 ^{14}C yr BP, summers were significantly warmer than today, and effective moisture increased markedly. To what extent did these climate changes affect the people in central Alaska?

Ecology of large herbivores. In order to assess how people would have been affected by changing climates, one needs to know how valued prey species would also have been affected. Limited faunal evidence indicate that bison (*Bison* cf. *priscus*) and elk (*Cervus elaphus*) were important food sources in the early Holocene (Powers and Hoffeecker, 1989; Yesner et al., 1992). By the mid-Holocene, however, caribou (*Rangifer tarandus*) become important (Betts, 1987), although at one site, bison were still hunted until the late Holocene (Yesner et al., 1992). Caribou remain important today and their bones are common in late prehistoric and protohistoric archaeological sites (Yesner, 1989; Betts, 1987; Shinkwin, 1979). Surprisingly, the archaeological record suggests moose (*Alces alces*) were not important, although they clearly are today and were in protohistoric times (Yesner, 1989; Guedon, 1974; Shinkwin and Case, 1984). Birds, fish, and small mammals also contributed to prehistoric diets, but since they would have been only indirectly affected by vegetation changes (fish), or poorly preserved (birds), or would have constituted comparatively little to the total annual diet (snowshoe hares), they are not discussed further.

During the full-glacial, bison were common in central Alaska, forming 1/3 of the "big three" triumvirate of mammoth, bison, and horse (Guthrie, 1968). The Pleistocene bison (*Bison priscus*) is larger than its presumed modern descendants (*B. bison bison*) (Guthrie, 1990a). Today, bison are no longer present in Alaska, but apparently vacated the north as recently as the late Holocene (Stephenson et al., in prep.).

An extensive analysis of plant remains preserved in *B. priscus* teeth and gut indicate that during the full glacial, *B. priscus* ate mainly grasses, with sedges, forbs, and woody plants coming in a very distant second (Guthrie, 1990a). Modern plains bison (*B. bison*) also prefer grasses (especially in the short grass prairie), although in marginal semidesert locations browse (prickly pear cacti) are consumed also. In Canada, woodland bison (*B. bison*

athabascaae) rely heavily on sedge, which contributes more than half of the graminoid diet; browse is strictly avoided (Reynolds, et al., 1982).

In sum, bison prefer grasses, but if necessary, will eat other plants. In the northern part of its range, modern bison live in wooded parklands of lodgepole pine (*Pinus contorta*) with extensive sedge and grass meadows (Yellowstone National Park). In northern Alberta, the woodland bison live in a boreal forest parkland with large sedge meadows, particularly near the Slave river. This indicates, that while bison are rightly considered a plains species, they also do well in wooded regions if sufficient openings with flourishing graminoids are also present (Reynolds, et al., 1982).

Elk (*Cervus elaphus*) are no longer present in Alaska, and the youngest fossil (recovered from the Yukon) dates to the mid-Holocene (Matthews, 1982).

Modern elk (found mainly in the western alpine North America) are combined grazers and browsers, changing their forage in response to seasonal abundance. During the winter, many populations subsist almost entirely on browse, such as oak, aspen, maple, sage, and willow. During the summer, most populations concentrate on herbs and forbs, although browse is not excluded (Peek, 1982; Hayden-Wing, 1979; Hobbs et al., 1979; Marcum, 1979).

In sum, elk have flexible food preferences and often migrate between wooded and open countryside. During calving, they tend to congregate near the timber/open ecotones, so that calves have good cover after birth (Peek, 1982).

Caribou (*Rangifer tarandus*) were present in Alaska during the full glacial, and are found today throughout much of the state (Guthrie, 1968; Matthews, 1982). Caribou are circumpolar, covering the Arctic and subarctic. Their range includes the boreal forest and tundra, with many populations wintering in the forest and migrating to the tundra during the summer.

During the winter, caribou feed mainly on lichen when it is abundant, although some green foods (bases of sedges, *Equisetum* stems, *Vaccinium vitis-idaea* leaves) are also eaten, possibly for their nitrogen content (Russell and Martell, 1984). Willow and *v. vitis-idaea* twigs are also eaten on occasion (Miller, 1982). In tundra areas, lichen may constitute less than half of the

winter diet, the remainder being mosses, grasses, and forbs. In forested areas, *Cladina* and other ground-growing lichen are preferred, although in deep snow, lichen growing on trees are also important.

While caribou are well adapted to snowy environments (Russell and Martell, 1984), they are sensitive to deep snow, especially snow with a crust. To reach the low-growing lichen, caribou must "crater" the snow with its foot. In forested settings, the caribou can reach the ground-lichen if the snow is less than about 75 cm; with deeper snow cover, the caribou switch to the tree-growing lichen or migrate to a less snowy area. In tundra and alpine areas, 60 cm of snow appears to be the maximum depth that they can effectively crater (Russell and Martell, 1984). When the snow develops a crust (a common event in maritime regions), caribou have extreme difficulties (Miller, 1982).

During the summer, most caribou migrate to tundra areas, where grasses, sedges, and forbs are a significant part of their diet, to the near exclusion of lichen (Miller, 1982).

In sum, caribou live either all year in the tundra, or migrate there during the summer. Their winter forage is dominantly lichen, although graminoids and mosses may also be important. During the summer, graminoids and forbs are their main diet. Caribou are sensitive to deep and/or crusted snow.

Even though the archaeological record suggests that moose were not important to people until relatively recently, they are included here because they may have appeared in Alaska only after the full glacial, and because they are an important feature of the modern boreal forest.

The modern moose (*Alces alces*) was apparently absent from Alaska during the full glacial. An extinct form of moose (*Cervalces latifrons*) was present during the last interstadial (about 28,000 to 40,000 yr BP), but no moose bones (either modern or extinct) have been recovered from full glacial deposits (Guthrie, 1990b; 1995). As yet, based on a small sample, the earliest *A. alces* bones from Alaska date to the early Holocene (Guthrie, 1990b).

Modern moose are boreal forest browsers. They prefer stands of riparian willows, but other browse (leaves or twigs from birch shrubs and trees, aspen saplings, and balsam poplar saplings) are also willingly eaten (Telfer, 1984;

Coady, 1982; Miquelle, et al., 1992; Van Ballenberg, 1992; Hjeljord et al., 1990). Moose also favor recently burned areas, where highly palatable browse (willow, poplar or aspen saplings) can be abundant (Loranger, et al., 1991). Alaskan moose apparently avoid conifer needles, although in eastern North America needles can form an important part of the winter diet (Telfer, 1984).

Moose are also aquatic grazers, possibly because aquatic plants contain needed minerals (Telfer, 1984). In some parts of Alaska, however, aquatics are not an important part of the summer diet (Van Ballenberghe, 1992).

In sum, moose are a boreal forest species, venturing into tundra environments only where riparian willow is abundant (D. Hopkins, pers. comm.). Willow, particularly in riparian settings, is their forage of choice, resulting in relatively high animal densities. However, moose will also eat other browse and they are generalists in their foraging strategy.

Potential Effects of Climate and Vegetation Change on People

12,000 to 11,000 ^{14}C yr BP. This period is marked by increasing warmth and effective moisture, causing dramatic changes in the vegetation.

The vegetation changes are marked by increasing plant densities and the widespread appearance of *Betula*. Willow was probably as abundant at this period as it is today. The vegetation was probably a mosaic, with the *Betula* and *Salix* occupying the mesic sites, while the graminoids occupied the xeric sites, such as south-facing slopes and ridge tops. The *Salix* was probably most abundant on floodplains and stream sides, while the *Betula* probably grew in moist, but poorly drained areas.

This transition to a shrub-dominated landscape would have affected the herbivores and the people. The earliest archaeological sites in Alaska (Mesa, Dry Creek, Walker Road, Broken Mammoth, Swan Point, and others [Figures 1.3 and 2.1]) (Kunz and Reanier, 1995; Powers and Hoffecker, 1989; Goebel, et al., 1991; Yesner, 1996; Holmes et al., 1996), date to this transition, and a possible causative link is intriguing. The spread of a shrub tundra would have made fuel more abundant, a particularly important feature for winter survival. I speculate that the ability of humans to travel across Beringia was limited by their ability to withstand the winter, and the spread of shrub

tundra may have allowed the people to over winter in formerly uninhabitable areas. That being said, how would this vegetation change affect the herbivores?

The moose populations would have been densest in river and stream valleys, where *Salix* was abundant, although the widespread shrub birch suggests they could have been found almost anywhere outside of the tundra zone. Elk, because it is a browser and grazer would have been found where both habitats were present, possibly near the birch/tundra ecotone or in openings in the shrub birch vegetation. Bison, a grazer, would have been locally restricted, concentrating in areas of open landscape, either in the tundra, or in areas where *Betula* was not common because the ground was too wet or too dry.

The concentration of the bison and elk into discrete grassy areas may have made it easier for the people to hunt them, because the game would have been in larger groups and at reasonably predictable locations. Faunal remains at the Broken Mammoth site (Figures 1.3 and 2.1) suggest elk and bison were the most common prey. This bluff-top site would have been an excellent location to spot game on the Tanana River floodplain, as well as on the grassy south-facing slopes above it (Yesner et al., 1992).

11,000 to 10,000 ^{14}C yr BP. The pollen records suggest that herbaceous taxa increased briefly during the Younger Dryas event. These taxa (grass and sage) probably occupied the most xeric sites while the mesic sites continued with *Betula* and *Salix*. The amount of vegetation change was quite small and would have been noticeable only on sensitive sites (such as ridge tops and south facing slopes), where water would have become limiting.

Faunal remains at Broken Mammoth suggest people were primarily hunting bison, and secondarily pursuing elk. Dall sheep (*Ovis dalli*) are also preserved at both Broken Mammoth and Dry Creek (Yesner et al., 1992; Guthrie, 1983).

I speculate that the small vegetation changes seen in the pollen records would have affected the people to the extent that they would have hunted in different localities for their favored prey. For example, bison might have become more abundant at lower elevations in especially arid areas, and elk

migrations might have shifted on the local scale to accommodate the new vegetation mosaic. The fundamental settlement pattern of seasonal residential camps with outlying overlook sites probably was unchanged, although the locations of the outlying sites would reflect the new herbivore distributions. That the Broken Mammoth and Swan Point sites (both hunting camps) (Yesner, 1996; Holmes et al., 1996) were apparently unoccupied at the beginning of the period supports this hypothesis.

Whether the increased aridity and cooler temperatures excluded people from the higher elevations or exposed locations is unclear. Dry Creek, while not a high elevation site, is exposed today to strong winds funneling through the Alaska Range, suggesting that during the YD, deteriorating climatic conditions could have lead to site abandonment. However, the inconsistent radiocarbon dates at Dry Creek (Figure 5.1) (Powers and Hoffecker, 1989; Bigelow and Powers, 1994) make it impossible to know exactly when the site was occupied during this period.

We have not, as yet discovered any high elevation sites that are firmly dated between 11,000 and 10,000 ^{14}C yr BP. Sites around the Tangle Lakes (Figure 2.1) apparently were not occupied until about 10,000 ^{14}C yr BP (West et al., 1996a; West et al., 1996b; West, 1981), suggesting the possible exclusion of humans from high elevations. However, the absence of sites is negative evidence and may be solely the result of faulty survey strategies.

10,000 to 8500 ^{14}C yr BP. During this period, *Populus* trees were more common on the landscape than before or since. They probably shared the riparian areas with the willow, possibly also occupying the warmer hill slopes, where moisture was not limiting.

Broken Mammoth was occupied at the beginning of this period. Faunal remains there are dominated by bison, with elk second in abundance.

Other sites from this time period have no faunal remains, but they are located usually in overlook locations indicating that we have only recovered the specialized sites related to subsistence strategies (Maxwell, 1987; Goebel and Bigelow, 1992; Bowers, 1980, Lively, n.d.). One group of sites in a slightly different setting are those at high elevations surrounding Tangle Lakes (West et al., 1996a; West et al., 1996b; West, 1981). These sites are associated with a

+30 m lake shoreline, formed when the lake outlet was dammed by a late Wisconsinan esker. People could have been fishing, although it is unlikely that these proglacial lakes were especially productive. More likely they were hunting caribou (and possibly bison or elk) at this high elevation locality.

The expansion of *Populus*, and at some localities, *Salix* during this period suggests good moose browse was abundant on the floodplains and stream valleys. Why are moose absent in archaeological sites from this period (and very rare up to the recent past)? Moose were present in the region, and it is probable that they traveled from Siberia to Alaska at about the same time as the people did (Guthrie, 1990b).

There are two possible explanations. First, the bison, elk, and caribou may have provided for all the necessary subsistence requirements, and these species, because they could be found in open areas, would have been easier to search and pursue than moose. Second, we may not have located the specialized moose hunting sites, as we have for the bison, elk, and caribou. Moose hunting sites would have been located in openings within the dense riparian thickets of willow and *Populus*, probably near running water. Today, these sites are deeply buried, or entirely destroyed. I speculate that the answer is a combination of the two possibilities. That is, the bison, elk, and caribou may have been preferred because they were easier to hunt, but that because of the realities of site preservation and visibility, we are missing a large part of the subsistence story.

8500 to 6000 ^{14}C yr BP. This period is marked by the appearance and spread, first of spruce, then of alder. By the end of the period, the primary taxa that constitute the modern boreal forest were present on the landscape, although this forest was significantly different from what we see today. The dominance by white spruce at the beginning of the period suggests the modern swampy landscape common today had not yet developed.

The archaeological record shows the continued occupation of some previous sites, although by the end of the period they are mostly unoccupied (Maxwell, 1987; Goebel and Bigelow, 1992; Yesner et al., 1992; Bowers, 1980, West, 1981). At Broken Mammoth, faunal remains are less abundant (with a few identifiable caribou and bison remains) because the bones in this cultural

horizon have been degraded by water leaching through the overlying acidic surface soil (Yesner et al, 1992).

The expansion of spruce forest marks a significant vegetation change. Presumably this resulted in contraction of the bison to higher elevations or other areas that are too dry or too wet for trees and shrubs. Modern (reintroduced) bison in Alaska apparently survive (but do not thrive) in limited areas where grass and sedge flourish, such as exposed, windy ridgetops and swampy areas (Guthrie, 1990a).

The effect on the people would be a need to radically change subsistence strategies. Even if bison were still common, they would have been harder to hunt, because of the blocking vegetation. The effect on caribou may have been positive, because if the forest was less swampy than today, it may have supported a lichen woodland similar to which is found in Canada; these sorts of open woodlands provide good winter forage. The spread of spruce may have limited the elk range, but it is unlikely, given their ability to live in coniferous forests today, that they went locally extinct.

With the spread of spruce and other arboreal species, overlook sites, particularly those at low elevations, may have become ineffective for spotting game. It is probably significant that all of the sites (all overlook) occupied at the start of the period, were unoccupied by about 7000 ^{14}C yr BP. As the forest gradually covered the landscape, hunting sites would have been located in areas of known game abundance (i.e., in stream valleys or on the floodplain, or in areas of patchy forest and graminoid vegetation), and not necessarily on easily recognized promontories.

6000 ^{14}C yr BP to the present. This period is a continuation of the previous period, with increased spruce and alder abundance on the landscape. There is some indication that the relative abundance of spruce, alder, and birch fluctuated, changing the composition of the boreal forest, but probably not its geographic extent. After about 2000 ^{14}C yr BP, spruce become less common on the landscape, possibly due to widespread paludification.

Elk were probably went locally extinct near the beginning of this period (the youngest elk bones [recovered from the Yukon] date to about 5000 ^{14}C yr BP [Matthews, 1982]), while the bison disappeared at the end of the period

(Stephenson et al., in prep.). Given elk's ability as a browser and grazer, it is a little surprising that it apparently died out first. One possible answer may be that as grassy patches within the forest diminished, the bison and elk came into increasing conflict. Today, where elk and bison are found together, the bison are usually dominant and force the elk to eat elsewhere (Reynolds, et al., 1982).

Archaeological sites during this period are located in the forest, along lakes or rivers, and above tree line. Faunal remains are scarce throughout much of the period, although wherever preserved, they indicate caribou were the favored large-game. Presumably moose were also hunted, but moose faunal remains are significantly less common than caribou (Yesner, 1989).

There is circumstantial evidence that widespread paludification in central Alaska did not begin until the latter half of the Holocene (Hu et al., 1996), perhaps not until after 2000 ^{14}C yr BP. Swamp development would have lead to the concentration and the final local extinction of bison, in addition to the concentration of moose. Prior to paludification, moose may have been disbursed more widely on the landscape, as swamps expanded, however, their preferred range may have been restricted to riparian settings. It may be that it was only after paludification that it was possible to reliably hunt moose, because they were more likely to be found in limited and reasonably predictable areas, such as along stream margins.

APPENDIX A—METHODS

Pollen Core Collection and Storage

Pollen cores were collected from rafts or through the ice using a modified Livingstone piston corer with a 6 cm barrel (Wright et al. 1984). The cores were extruded in the field into semi-circular PVC pipes, covered with another semi-circular pipe, and taped shut. The vegetation surrounding each lake was noted, and samples of the aquatic flora were collected for further identification in the lab. The lake basins were mapped in detail, using either a Global Positioning Satellite (GPS), or a tape and lead.

Upon return to the laboratory, the cores were refrigerated at 5°C. Prior to analysis, each core was split in half lengthwise, with one halve saved for archiving. The work half of the core was cleaned, described, and in most cases, photographed prior to sampling.

Pollen Preparation and Analytical Methods

1 cc or 2 cc volumetric pollen samples were collected from the cleaned core surface. Prior to processing, two to three tablets of known quantities of *Lycopodium clavatum* spores were added to each sample for calculating pollen influx (Stockmarr, 1972; Davis, 1966). Processing followed the methods of Faegri and Iversen (1989), that is, dilute acid and base washes to remove carbonates and humates, acetolysis to remove cellulose, concentrated hydrofluoric acid to remove silts, staining with safranin, and finally, mounting in silicon oil. Pollen taxa were identified by comparison with pollen reference collections and with various pollen keys, such as Faegri and Iversen (1989), Moore, et al. (1991), McAndrews et al. (1973) and Moriya (nd).

Pollen samples were counted until 300 pollen grains from trees, shrubs and herbs/forbs combined were encountered; this forms the basic pollen sum. To calculate percentages, the percentage of trees, shrubs and herbs/forbs was based on the basic pollen sum. Percentages of spores is based on the basic pollen sum + spores, aquatics is based on basic pollen sum + aquatics (excluding *Pediastrum*), and *Pediastrum* is based on basic pollen sum + *Pediastrum*.

Betula grains were not measured to separate *Betula papyrifera* (tree birch) from *Betula glandulosa/nana* (shrub/dwarf birch), because these species hybridize today. Analyses of modern *Betula* grains demonstrate that grain measurements do not accurately separate these two groups (Edwards et al., 1991).

Picea grains were qualitatively separated into three groups, *Picea* undiff., *Picea glauca*-type, and *Picea mariana*-type, using the methods outlined in Hanson and Engstrom et al. (1985).

Macrofossil Methods

Macrofossil analysis was conducted on one core. Slices two cm thick were taken from the work half of the core. The volume of the slices was measured by displacement in a 100 cc measuring cylinder. The sample was then transferred to a 180 μ screen, and sieved gently with a water spray. The portion passing the screen was saved for possible future pollen analysis. The portion caught in the screen was transferred to a ziploc bag for later identification. The seed and plant fragments were identified using a dissecting microscope with magnification ranging between 10x and 50x. Identifications of the macrofossil remains were made based on discussions with Dr. Hilary Birks, and comparison with a macrofossil reference collection.

AMS Samples

Macrofossils for AMS analysis were collected by gently sieving core slices at 180 μ . The macrofossils were then dried overnight at 80°C and weighed.

Several samples of *Picea* pollen were concentrated for AMS analysis. The pollen was concentrated using modified methods of Brown et al. (1989). A cleaned slice from one half of the core was sieved at 250 microns; the portion passing through the sieve was washed with weak acid and base solutions to remove humates and carbonates. This fraction was then suspended in a heavy liquid solution of inorganic, non-toxic, water-based sodium polytungstate at a specific gravity of 2.0. The sample was then spun down for 3 minutes at 2000 rpm. The portion floating in the centrifuge tube was poured off into a buchner funnel with a 2 micron glass fiber filter inside. A

weak vacuum extracted the remaining heavy liquid from the sample in the buchner funnel.

The glass fiber filter with the sample on it was transferred to a 50 ml centrifuge tube; concentrated HF was added to the tube to dissolve the filter. The sample was then spun to decant the HF. A weak acid (10% HCl) was added and the sample was placed in a hot water bath (80° C) for 10 minutes. This removes flouride precipitates resulting from the HF wash. The sample was then bleached in a 50% solution of household bleach for five minutes and subsequently washed three times in distilled water. The sample was then sieved at 180 micron, 45 micron, and 20 micron sieve sizes. The *Picea* pollen is concentrated in the 45-180 micron fraction. The ≥ 180 micron fraction contained abundant unidentifiable organic debris and *Pediastrum* nets. The 45-20 micron and < 20 micron fraction contained some *Betula* and *Alnus* pollen, but also abundant small fragments of amorphous organic matter. The 45-180 micron fraction, while containing abundant *Picea*, also contained other organic matter. To improve the quality of this fraction, it was resuspended in a new solution of Sodium polytungstate with a specific gravity of 1.8.

At specific gravity 1.8, the sample continues to float. Judicious dilution with a pipette causes the heavier organic matter to sink, while the *Picea* remains floating. This organic matter is removed with a pipette. The sample is diluted more, so the *Picea* begins to sink, but the lighter organic matter continues to float. As the *Picea* sinks it is removed with a pipette and transferred to a different centrifuge tube. The *Picea* can be recognized by the naked eye because it is a light yellow color with a slight granular texture. The heavier organic matter is brown or black, while the lighter organic matter is white, and often appears fluffy in the centrifuge tube. This step can be tedious and requires some practice to recognize the *Picea*, but the final sample is very clean, with a minimum of pollen being lost with the organic matter. The sample is then washed in distilled water, transferred to a weighed vial, and dried in an oven at 80° C for 12 hours. The vial can then be reweighed on a balance (to least four decimal places) and the sample weight calculated.

The advantages of this method is it allows the researcher to date any location in the core where there is *Picea* pollen. Smaller pollen grains have

been dated (Regnell, 1992), but my experience at Windmill and other lakes indicates that *Betula* pollen, because of its smaller size and weight, is harder to separate (either with sieves or heavy liquids) from unwanted organic matter. In addition, because of *Betula*'s lighter weight, larger slices of the core are required to achieve a clean sample large enough for AMS analysis.

The disadvantages of this method is it is time-consuming, with ample opportunity to contaminate the samples, and expensive. It takes two to three days to process a batch of four samples. One reason the method is slow is every lake is different, so some samples may need several KOH washes to remove the humates, while others need only one. In addition, the volume and kinds of organic matter affect one's ability to reliably separate the *Picea* pollen from other organic matter. For example, *Pediastrum* nets are often the same size, and only slightly lighter than the *Picea*, while *Ceratophyllum* spines are often the same size and weight as the *Picea*. Samples that had limited amounts of these organics were the easiest to process.

As the sample becomes cleaner, particularly in the final steps, it becomes more and more important not to contaminate it with modern carbon. The samples must be kept covered whenever they are not being worked on. One should be wary of dropping bits of dust, hairs, etc. into the sample while pipetting out contaminants.

Finally, the method is expensive because new Sodium polytungstate must be used for each sample. Sodium polytungstate costs about \$98/pound, this works out to about \$10 per sample. With repeated use, the heavy liquid turns from colorless to a light brown color, possibly because it is picking up dissolved organics in the sample. One can reuse the same heavy liquid for each individual sample, but after the initial separation, the heavy liquid is quite dilute, requiring significant evaporation in the oven before it can be reused.

Loss on Ignition and Magnetic Susceptibility

For Loss on Ignition, samples were dried for 20-24 hrs in an 80°C oven, weighted to .01 g accuracy and burned in a muffle furnace. For estimates of organic carbon, the samples were burned at 500-550° C for two hours. The

samples were then reweighed and returned to the furnace for an additional two hour burn at approximately 850° C to estimate carbonate content. After the final burn, the samples were again weighed. The percent weight loss was calculated as a percentage of the original dry weight.

Magnetic susceptibility was measured on the whole core using a Bartington susceptibility meter with a ring. All cores were measure in continuous 2 cm intervals.

Carbon and Nitrogen Measurements

Carbon/Nitrogen concentration and isotope analysis was conducted on samples collected at varying intervals for each core. The size of the sample depended on the amount of organic carbon, samples with higher organic carbon were smaller than those with low organic carbon. The samples were put in sterile centrifuge tubes and washed with a 1N HCl solution to remove carbonates. After the sample stopped fizzing (most samples had no reaction, and those that did were washed twice in HCl), the sample was then washed three times in distilled water and dried in the centrifuge tube at 80° C for 24 hours. The samples were then analyzed on a mass spectrometer.

Carbon and nitrogen fluxes ($\text{g}/\text{cm}^2/\text{yr}$) were calculated by transforming the N and C percentage data into proportions, then multiplying by the dry bulk density (g/cm^3) and sedimentation rate (cm/yr). The dry bulk density was measured by sampling with a 1 cm^3 volumetric sampler, and weighing the sample after drying 24 hours at 80°C.

Some samples, however, did not have bulk density measurements. In these cases (the samples retrieved between 900 and 1000 cm in Dune A), the bulk density was reconstructed by plotting a regression between water content and bulk density for lower samples in the core. This regression equation was then used to estimate the bulk density of the upper samples (which did have water content measurements).

Statistical Analyses

Statistical analyses on the pollen data include zoning, rate of change, and ordination. Except where otherwise specified, all of these analyses were

undertaken on percentage data of the terrestrial taxa with a maximum abundance of 2%, and which occur more than 5 times in a given core. In practice, this means the arboreal and shrubby taxa, plus the most common of the herbaceous taxa.

Zoning. Computerized zoning of the pollen diagrams was accomplished using the computer program Zone (ver. 1.2, Birks and Line, 1991). Various agglomerative and divisive techniques were used to help identify the most consistent pollen zones. Agglomerative methods (constrained incremental sum of the squares [CONISS] and constrained single linkage [CONSLINK]) begin with each sample in its own cluster; the variance of each cluster increases as samples are added to existing clusters. In contrast, divisive methods (optimal sum of squares partition (OPTSSQ), split least squares [SPLITLSQ], split information statistic [SPLITINF]) begin with the entire pollen core as one cluster, which is divided at the point where adjacent samples are most dissimilar. In optimal division (OPTSSQ), each successive split is unrelated to previous splits (no hierarchy is implied). In binary division (SPLITLSQ and SPLITINF), each successive split is made within existing zones, creating a hierarchy (Birks and Gordon, 1985; Bennett, 1996).

One difficulty with using computer algorithms for zoning a diagram is deciding at what level the zones become trivial. Bennett (1996) addresses this problem by comparing various characteristics of the zones in a stratigraphic dataset with the zones in the same dataset that have been stratigraphically randomized. With an agglomerative zoning technique (CONISS), one compares the increase of the variance within the zones as the clusters are joined together. When the within-zone variance (as a proportion of the running sum of total explained variance) of the actual dataset diverges from the randomized dataset, then any zones with a greater explained variance are considered significant, while those with lower explained variance are considered trivial.

With divisive techniques, one compares the change in the residual (unexplained) variance after each partition. When the change in the unexplained variance converges with the variance change in the randomized

data, then the partitions up to that point are significant, while those coming later are trivial.

For this analysis, the datasets were randomized by assigning a computer-generated random number to each sample, and then sorting the samples according to their assigned random number. This was repeated three times to assure that the data were well mixed.

Rate of change. Rate of change analysis was accomplished by calculating dissimilarity coefficients (DC) between adjacent samples and standardizing to a unit time interval. The various dissimilarity coefficients are noted in Table A.1.

Table A.1: Dissimilarity coefficients (Prentice, 1980).

Name	Formula	Comments
Euclidian distance	$d_{ij}=[\sum_k(p_{ik}-p_{jk})^2]^{1/2}$	simple DC
Standardized Euclidian distance	$d_{ij}=[\sum_k(p_{ik}-p_{jk}/s_k)^2]^{1/2}$	equal weight to common and rare taxa
Chord distance	$d_{ij}=[\sum_k(p_{ik}^{1/2}-p_{jk}^{1/2})^2]^{1/2}$	emphasize signal to noise
Chi-squared 2	$d_{ij}=[\sum_k(p_{ik}-p_{jk})^2/T_k]^{1/2}$	emphasize signal to noise

Where: p_{ik} =proportion of taxon k for sample i
 p_{jk} =proportion of taxon k for sample j
 s_k =standard deviation of taxon k for all samples
 T_k =total proportion of taxon k for all samples

The choice of DC depends on the individual pollen core. Simple DCs all but ignore minor taxa, while standardized DCs give minor and major taxa equal weight. Preference is given to the DC which maximizes signal to noise (Prentice, 1980; Overpeck et al., 1985; Bennett and Humphry, 1995). The *rate of change* is the DC divided by the time step between the samples. This statistic, however, is highly sensitive to the sample interval, because small intervals automatically lead to high rates of change. This is alleviated by calculating the DCs on a dataset that has been interpolated to a unit sample interval. For the purposes of this analysis, 100 yr was chosen as the standard because it is close to the mean sample interval for all cores (ranging between

about 140 and 200 years among the Windmill, Dune, and Birch lake cores), and it allows comparison between the different cores.

Ordination. Ordination of the pollen data was conducted using the computer program CANOCO (ver. 3.10) (Ter Braak, 1988) on square-root transformed data percentage because it down-weights the major taxa relative to the minor taxa. A Detrended Correspondence Analysis (DCA) was carried out first to determine the gradient length of the first axis. If the gradient is less than 4 standard deviation units, then Principle Components Analysis (PCA) is an appropriate ordination technique (Ter Braak, 1986; Ter Braak and Prentice, 1988). PCA is a multiple linear regression of the data, using a least squares fit.

APPENDIX B— ISOTOPES

Carbon and Nitrogen Stable Isotopes

Isotopes are calculated with reference to a universal standard. For ^{13}C the standard is the PeeDee belemnite, for ^{15}N the standard is the atmosphere. The conventional notation is:

$$\delta X(\text{‰}) = ((\delta X_{\text{sample}} - \delta X_{\text{standard}}) / \delta X_{\text{standard}}) 1000,$$

where δX is the ratio of the relevant isotope, that is, $^{13}\text{C}/^{12}\text{C}$ and $^{15}\text{N}/^{14}\text{N}$

(Peterson and Fry, 1987; Ehrlinger and Rundel, 1988).

Both carbon and nitrogen stable isotopes can provide information about the lake and its watershed. The difficulty in all stable isotopes they are affected by a myriad of variables (Peterson and Fry, 1987). If all the source material is used in the reaction, no net fractionation takes place, and the isotopic signal of the product is the same as the source. If only some of the source is used, then fractionation (discrimination against the heavy isotope, that is against ^{15}N and ^{13}C) typically takes place during photosynthesis. The amount of fractionation is thus dependent on the amount of source material, as well as growth rate.

The second variable is the isotopic signal of the source material itself. For limnological studies this means the difference between different kinds allochthonous and autochthonous inputs. Autochthonous inputs include decayed organic matter from the watershed, in addition to sources from rain, surface run-off, and ground water. Allochthonous inputs are materials formed within the lake itself.

Below are tables listing some of the factors affecting $\delta^{13}\text{C}$ values (Tables B.1 and B.2).

Table B.1: $\delta^{13}\text{C}$ as related to source carbon.

Factor	$\delta^{13}\text{C}\text{‰}$ rel. to PDB	Reference
Source: Atmosphere	-7‰	Peterson and Fry (1987)
Source: C3 plants	-33‰ to -22‰	Bender (1971)
Source: C4 plants	-20‰ to -10‰	Bender (1971)
Source: Bicarbonate	0‰ to +2‰	Deuser and Degens (1967)

Table B.2: $\delta^{13}\text{C}$ as related to environmental variables.

Factor	Effect on $\delta^{13}\text{C}$ (other variables held constant)	Reference
Temperature	+2.4‰ from 0°C to 30°C (aqueous bicarbonate) ca. +0.25‰/°C (marine diatoms) +0.35‰/°C (marine plankton)	Deuser and Degens (1967) Degens et al. (1968) Fontugne and Duplessy (1981)
CO ₂ (aqu)	Decrease $\delta^{13}\text{C}$ with increase pCO ₂	Rau (1994)
Lake productivity	Increase $\delta^{13}\text{C}$ with increase productivity, because of CO ₂ drawdown during photosynthesis	Gu et al., (1996)

Lakes have several different carbon sources, these include CO₂ from the atmosphere, rain, surface run-off, and ground water. Particulate organic matter derives either from the surrounding watershed, or is manufactured within the lake. Inorganic carbonates (CaCO₃ or HCO₃⁻) can originate from the bedrock, although the bicarbonate is also produced by photosynthesis. All of these factors affect the $\delta^{13}\text{C}$ signature of an organism. Lacustrine autotrophs use either dissolved CO₂ or HCO₃⁻ for photosynthesis. Most species use dissolved CO₂, although a few use bicarbonate (Keeley and Sundquist, 1992).

The nitrogen isotope signal is difficult to interpret, primarily because the nitrogen cycle in both terrestrial and aquatic settings is very complex. Nitrogen-fixing bacteria have a $\delta^{15}\text{N}$ value close to 0‰ (Handley and Raven, 1992). Terrestrial plants in non-coastal and non-agricultural settings have a $\delta^{15}\text{N}$ have between -9‰ to +8‰ with a mode at -1‰ (France, 1995). $\delta^{15}\text{N}$ does not separate between C3 and C4 plants, and only clearly separate nitrogen-fixers from non-fixers in large studies (nitrogen-fixers having approximately 2‰ less ¹⁵N than the non-fixers) (Heaton, 1987; Virginia and Delwiche, 1982).

Nitrogen inputs into a lake are primarily nitrogen gas (N_2), nitrate (NO_3^-), and ammonium (NH_4^+). All three of these molecules come from the atmosphere, surrounding watershed, or are manufactured within the lake itself. Relatively few studies have been undertaken that examine the $\delta^{15}\text{N}$ of the various nitrogen fractions within a lake. One study, however, of a subarctic lake is directly applicable to this research. Gu and Alexander (1993a) noted two facts:

- 1) The $\delta^{15}\text{N}$ of nitrogen-fixing blue green algae was significantly less than for green algae (means are 1.0‰ and 6.6‰, respectively).
- 2) The $\delta^{15}\text{N}$ of dissolved nitrate and ammonium were not significantly different from each other, having a range between 5.6‰ and 7.1‰.

In lacustrine sediments, nitrogen isotopes are also useful for measuring allochthonous/ autochthonous inputs (France, 1995). France (1995) noted that the $\delta^{15}\text{N}$ became progressively heavier from terrestrial, to freshwater, to marine environments. Despite significant overlap in the $\delta^{15}\text{N}$ signals, terrestrial plants have a modal value of about -1‰, freshwater plankton and macrophytes have a mode about +3‰, and marine plankton have a mode about +6‰.

Finally, $\delta^{15}\text{N}$ also increases with trophic level, so that lacustrine zooplankton and fish have more $\delta^{15}\text{N}$ than their phytoplankton lunches (Estep and Vigg, 1985).

APPENDIX C--BONANZA BLUFF

Introduction

Pollen analysis at Birch and Windmill lakes suggest the vegetation between 14,000 and 12,000 ^{14}C yr BP was dominated by grasses, sedges, and *Artemisia*. Analog studies comparing fossil pollen with modern pollen in various vegetation zones indicate that while the grass-sedge-*Artemisia* pollen assemblage has no close modern analogues, this assemblage is most similar to the pollen spectra from high Arctic and fell field vegetation (Anderson et al., 1989; Cwynar, 1982). The main feature which prevents a close similarity between the fossil and modern pollen is the abundant *Artemisia* in the fossil assemblage. Today, *Artemisia* is most common in the prairies of southern Canada and the northern United States, prompting some researchers to suggest a steppic analog for the fossil assemblage (Guthrie, 1990a). *Artemisia* is not common in central Alaska today; it is limited to alpine settings, or at low elevations, to warm, well-drained sites such as riverine gravel bars and south-facing slopes (Hultén, 1968; Viereck et al., 1992). Are the sites in central Alaska that have abundant *Artemisia* today good analogues for pollen assemblages from 14,000 to 12,000 ^{14}C yr BP?

To address this question, modern pollen was collected from one south-facing slope to compare with the pollen assemblages from Windmill and Birch lakes. The modern pollen was collected in traps placed on the slope, and as a result, the data are only somewhat comparable to the data from lake cores. Pollen in lakes represent a mixing of local and regional pollen (Jacobson and Bradshaw, 1981). In contrast, pollen traps contain abundant local pollen, which can be highly variable, even over short distances. In addition, pollen traps also collect pollen from insect-pollinated taxa (i.e., forbs), while lakes, on the other hand, are dominated by wind-pollinated taxa (Faegri and Inversen, 1989). As a result, comparisons of the pollen data between these two types of collections must be made carefully, always keeping in mind their taphonomic differences.

Study site

Bonanza Bluff (informal name) is located approximately 30 km southwest of Fairbanks (Figure 2.1). The bluff is located on the southern slope of a hill system situated on the north side of the Tanana River. Paleozoic or Precambrian Birch Creek schist forms the hills, which are mantled with one to several meters of loess (Péwé et al., 1966). Depending on drainage and aspect, the south-facing slopes of the hills are vegetated by mixed stands of *B. papyrifera*, *Picea* (usually *P. glauca*), and *P. tremuloides*, with an understory of *Rosa acicularis*, *Viburnum edule*, and *Shepherdia canadensis*. Bonanza Bluff is located on the lower elevations of the south-facing slopes, where trees are absent, and the vegetation is dominated by *Artemisia*, grasses and forbs.

Due to the combined effects of aspect and high slope angle, Bonanza Bluff is very warm and arid today. Winter snow accumulation is low because winter winds remove it, and high spring insolation melts the snow quickly. Spring and summer air temperatures are significantly warmer than on the less steep wooded slopes, resulting in high evaporation. Calcium carbonate accumulations are present in the soil profile on the bluff, indicating an arid environment in which rainwater is unable to completely leach carbonates out of the loess.

At the bluff top, near the border with the trees, shrubs dominate. *Rosa acicularis*, *Shepherdia canadensis*, *Spirea*, and *Arctostaphylos uva-ursi* are common. *Zygadenus elegans* is also present (but not common) in this area. Further down the slope, graminoids and forbs, such as *Poaceae*, *Artemisia frigida*, *Pulsatilla patens*, *Antennaria* c.f. *rosea*, *Gallium borealis*, and *Erigeron* spp. become more common. Two juniper bushes (*J. communis*) (both male) are present on the middle to lower elevations of the bluff. Trees (*P. tremuloides*) are rare, and are limited to locations with a shallower slope angle, or where the bluff face does not face due south. The distribution of taxa is somewhat variable on Bonanza Bluff, probably due to micro-habitat variations in soil moisture, because the more moisture-demanding taxa are found in areas where the slope angle decreases and evaporation presumably is less intense.

Methods

Ten pollen traps similar to Tauber's design (1974) were constructed of 13 cm-long pieces of 10 cm diameter plastic pipe and capped by an aerodynamic wooden lid with a 6.5 cm-diameter hole. The hole in the lid was covered with chicken wire to prevent rodents from falling into the traps (one rodent, did, however, succeed in killing itself). Several of the traps accumulated abundant grasshopper carcasses, indicating a finer mesh wire would have been more appropriate. The traps were filled 1/4 to 1/3 full with an approximately 50/50 mixture of propylene glycol (antifreeze) and ethanol. Antifreeze prevents the traps from drying out, while the ethanol serves as a fungicide.

No effort was made to place the traps randomly at Bonanza Bluff. Instead, they were located to sample the widest variety of herbaceous vegetation. The traps were placed on April 30, 1995 (*Pulsatilla patens* had already started to bloom), and retrieved on September 4, 1995; thus the pollen accumulation is from only one season, and interannual variability has not been assessed. At each sampling locality, the trap was buried about 10 cm into the soil to prevent tipping. The vegetation at each locality was also rated on a Braun-Blanquet scale for 1 m and 2 m radii around the trap. Every two to three weeks the traps were replenished with the antifreeze/ethanol mixture, but the liquid in each trap was collected only when the traps were retrieved in September. One trap (#9) fell over during the course of the summer, so it was removed from the bluff, and not included in the analysis.

In the laboratory, the liquid from the traps was centrifuged to concentrate the pollen, and washed several times with water. Thereafter the processing followed the methods of Faegri and Iversen (1989). Pollen samples were counted until at least 300 terrestrial grains were identified. Some samples (traps 5, 6, 7, 8, 10) were overwhelmed by a single taxon, these samples were counted until the pollen sum excluding the abundant taxon reached 300.

Results

Pollen spectra. Pollen frequencies from the traps were calculated in a number of different manners. To compare the pollen in the traps with the late Holocene pollen in the pollen cores, all terrestrial taxa except *Hedysarum*-

type were included in the pollen sum (Figure C.1). *Hedysarum*-type is present in overwhelming quantities in one trap, while it is negligible in all other traps, indicating that random differences in pollen deposition significantly affected this taxon.

Picea (mainly *P. glauca*) pollen varies between 2% and 25% , while *Betula* and *Alnus* range between about 5% and 30%. *Juniperus* is usually low, but reaches about 12% in trap 2. *Salix* is highest in traps 7, 8, and 10, with maximum values of 73%. *Populus*, *Shepherdia*, and *Ericales* are low, with maximum values between 2% and 5%. *Artemisia* is common, with a maximum value in trap 6 of 64%, in other traps this taxon ranges between 1% and 22%. *Cyperaceae* and *Poaceae* are not common in the pollen traps, with maximum values of 2% and 5%, respectively. *P. patens* is present in all traps, although most abundant in trap 4, where the frequency is 12%. *Rosa*-type pollen (presumably *R. acicularis*) is over-represented in trap 5, with a value of 55%, in other traps it ranges between <1% to 9%. *Solidago*-type pollen, produced by the *Erigeron* and *Antennaria*, is present in all traps; it reaches maximum values in trap 4 at 13%.

To compare the Bonanza Bluff pollen spectra with fossil samples from Birch and Windmill lakes, the modern pollen was recalculated excluding the arboreal/shrubby taxa which were not abundant on the landscape 9500 to 14,000 ^{14}C yr BP (Figure C.2). Thus *Picea*, *Alnus*, and *Shepherdia* were excluded from the pollen sum. *Hedysarum*-type was also excluded. When the frequencies are recalculated, *Betula* and *Salix* combined account for approximately 10% to 95% of the spectrum. *Artemisia* and *Rosa*-type reach maximum values of 70% and 77%, respectively. *Populus* and *Ericales* have maximum frequencies of 3% and 7%, while *Cyperaceae* and *Poaceae* have maximum frequencies of 2% and 6%.

Ordination. To compare the Bonanza Bluff pollen spectra with the Birch and Windmill fossil spectra, all three datasets were ordinated together. The first axis gradient length of the square root transformed pollen spectra is about 2 standard deviations units, indicating PCA is an appropriate ordination technique for this dataset.

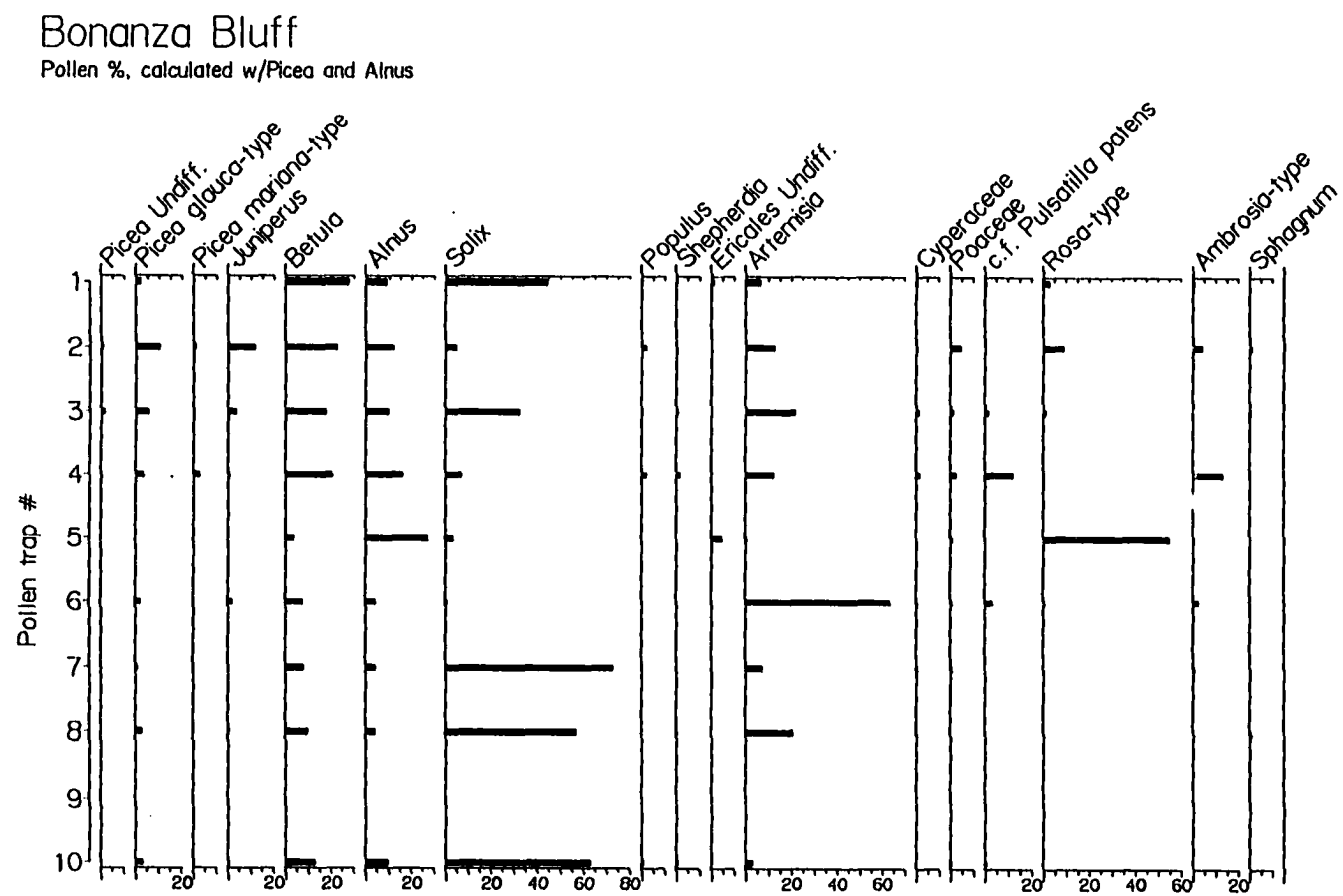


Figure C.1: Bonanza Bluff pollen percent (calculated with *Picea* and *Alnus*).

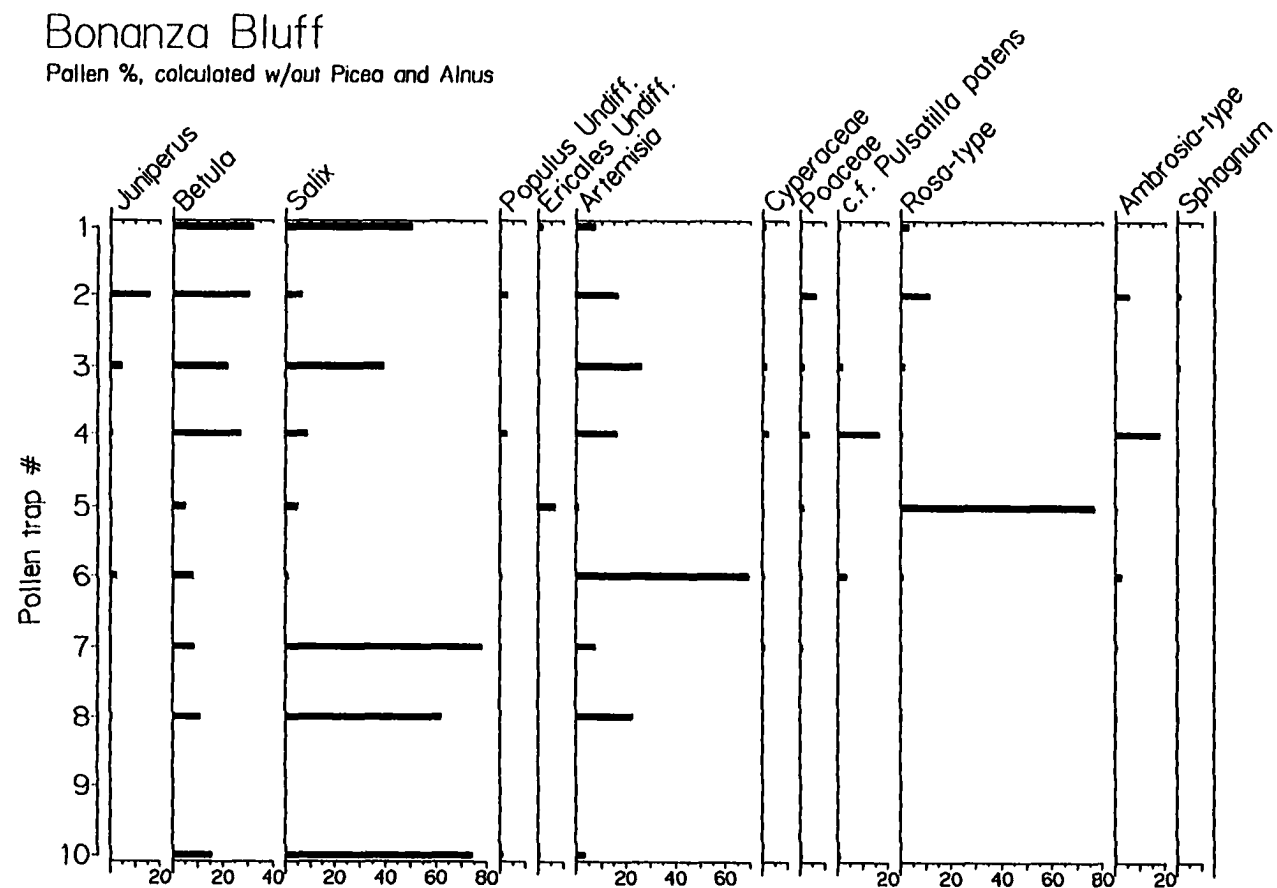


Figure C.2: Bonanza Bluff pollen percent (calculated without *Picea* and *Alnus*).

PCA was conducted on all species with a maximum abundance $\geq 2\%$, and in the case of the lacustrine samples, which also occurred five times or more in each sequence (Table C.1).

Table C.1: Taxa included in ordination with complete lacustrine cores.

<i>Picea</i>
<i>Juniperus</i>
<i>Betula</i>
<i>Alnus</i>
<i>Salix</i>
<i>Populus</i>
<i>Ericales</i>
<i>Artemisia</i>
Cyperaceae
Poaceae
Ranunculaceae, <i>Anemone</i> -type
Ranunculaceae, <i>Aconitum</i> -type
<i>Pulsatilla</i>
Fabaceae, <i>Astragalus</i> -type
Rosaceae, undiff.
<i>Rosa</i> -type
<i>Sanguisorba</i>
Caryophyllaceae
Brassicaceae
Asteraceae, <i>Solidago</i> -type

In the first PCA analysis, the Bonanza Bluff pollen frequencies include the modern arboreal taxa. Axes 1 and 2 respectively explained 52% and 23% of the variation in the dataset. A biplot of this analysis clearly separates the main pollen zones in the cores both from each other and from the modern taxa (Figure C.3). Because the Bonanza pollen frequencies were calculated with the modern arboreal taxa, the sample scores are intermediate between birch zone and herb zone taxa on the second axis, but closer to spruce zone samples on the first axis. Looking at the sample scores, it is clear that some Bonanza samples (traps 1, 3, 10) are strongly affected by those taxa which occur only in these samples, that is *Pulsatilla*, *Rosa*-type, and *Aconitum*. The traps with the highest frequencies of *Salix*, *Artemisia*, *Solidago*-type, and *Caryophyllaceae* (traps 6, 7, 8), are those closest to the herb zone samples. In contrast, the traps with the highest *Picea* and *Alnus* frequencies (traps 3, 4, 5) are those closest to the spruce zone samples.

Bonanza Bluff, Windmill Lake, & Birch Lake

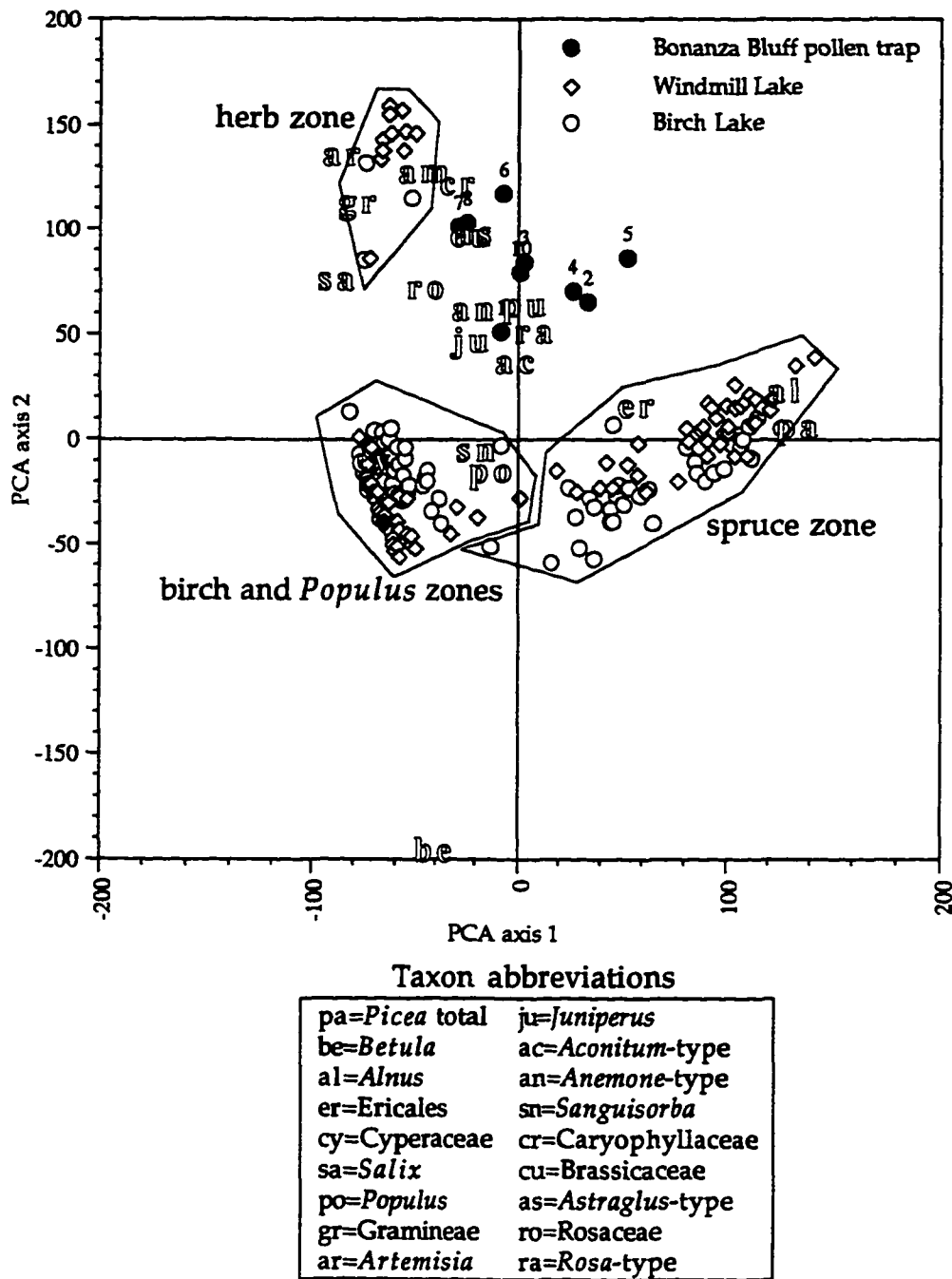


Figure C.3: Ordination of Bonanza Bluff, Birch Lake, and Windmill Lake pollen frequencies. Bonanza pollen frequencies calculated with *Picea*, *Alnus*, and *Shepherdia*.

Species scores for the first axis indicate *Picea*, *Alnus*, and *Ericales* have positive loadings, while *Juniperus*, *Betula*, *Salix* and the most of the herbaceous taxa have negative loadings. On the second axis, *Picea* and *Ericales* are near zero, while *Juniperus*, *Alnus*, *Salix*, and most of the herbaceous taxa have strong positive loadings. *Betula* has a strong negative loading on this axis (Table C.2).

Table C.2: Species loadings on axes 1 and 2.

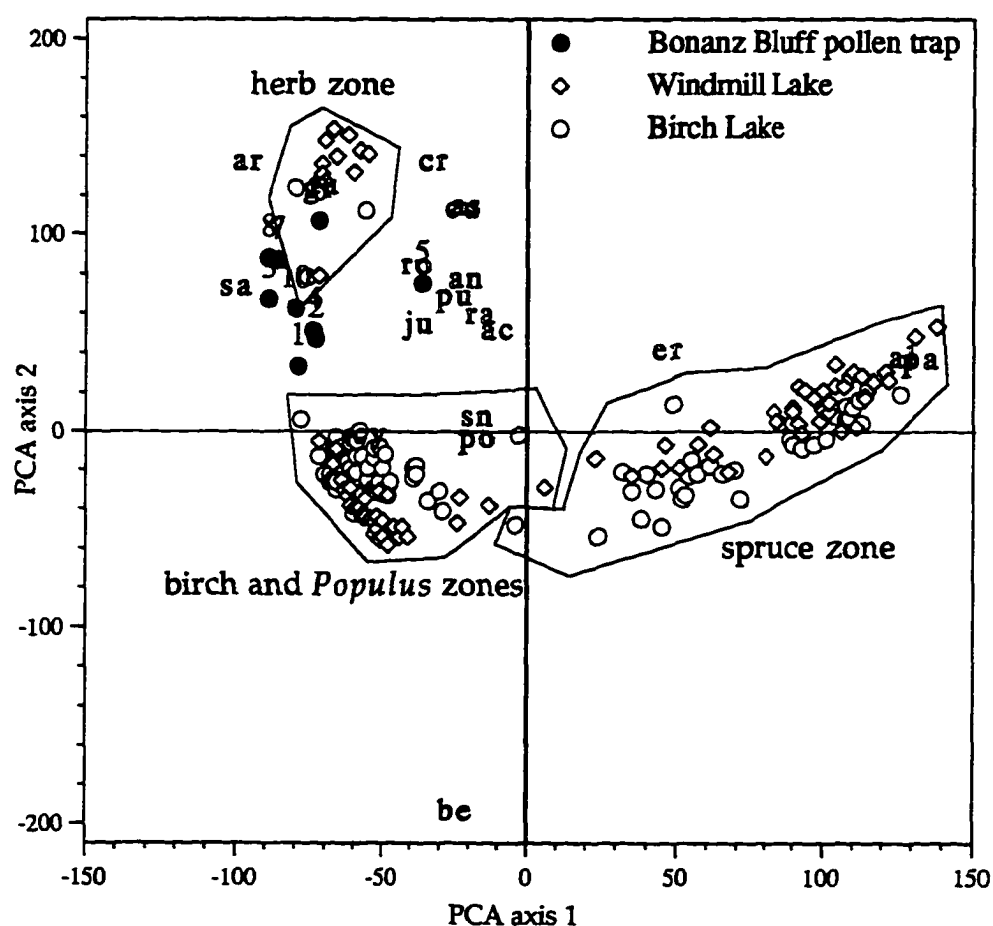
Species	Axis 1	Axis 2
<i>Picea</i> summed	134	6
<i>Juniperus</i>	-22	47
<i>Betula</i>	-40	-196
<i>Alnus</i>	128	23
<i>Salix</i>	-88	76
<i>Populus</i>	-12	-15
<i>Ericales</i>	58	16
<i>Artemisia</i>	-85	136
Cyperaceae	-70	-10
Poaceae	-77	113
<i>Sanguisorba</i>	-19	-5
<i>Aconitum</i> -type (Bonanza only)	-1	37
<i>Pulsatilla</i> (Bonanza only)	3	64
<i>Rosa</i> -type (Bonanza only)	8	51
Other herbs/forbs (mean)	-30	97

A second Principle Components Analysis excluding modern arboreal taxa (*Picea* and *Alnus*) was conducted on Bonanza pollen frequencies. In the new ordination, the sample scores for the Birch and Windmill samples do not change (Figure C.4). The primary effect is to move the Bonanza samples so they lie between the herb zone and birch zone fossil samples. The species scores do not change markedly, although the taxa that are found only in the Bonanza samples tend to have lower axis 1 loadings.

Discussion

Comparisons with fossil pollen. Ordinations of the Bonanza Bluff pollen with the fossil pollen suggest the modern spectra are somewhat different from the fossil spectra. The placement of the Bonanza samples

Bonanza Bluff, Windmill Lake, & Birch Lake



Taxon abbreviations

pa= <i>Picea</i> total	ju= <i>Juniperus</i>
be= <i>Betula</i>	ac= <i>Aconitum</i> -type
al= <i>Alnus</i>	an= <i>Anemone</i> -type
er= <i>Ericales</i>	sn= <i>Sanguisorba</i>
cy= <i>Cyperaceae</i>	cr= <i>Caryophyllaceae</i>
sa= <i>Salix</i>	cu= <i>Brassicaceae</i>
po= <i>Populus</i>	as= <i>Astragalus</i> -type
gr= <i>Gramineae</i>	ro= <i>Rosaceae</i>
ar= <i>Artemisia</i>	ra= <i>Rosa</i> -type

Figure C.4: Ordination of Bonanza Bluff, Birch Lake, and Windmill Lake pollen frequencies. Bonanza pollen frequencies calculated without *Picea*, *Alnus*, and *Shepherdia*.

between the spruce zone and herb zone fossil samples suggests Bonanza samples reflect climatic conditions both drier and colder than today, but not as cold and dry as the herb zone (Figure C.3). This interpretation is ironic because these localities are clearly abnormally warm today. This inconsistency underscores the problem with associating climate variables with vegetation. The Bonanza dataset appears cold partly because the locality is limited more by moisture than by temperature. Vegetation in cold dry places may not be all that different from warm dry places, because both are treeless, and may have abundant *Artemisia* and grasses.

Whether the Bonanza Bluff-type vegetation is a reasonable analog for the herb zone is difficult to say. The PCA suggests that Bonanza Bluff is most similar to the late herb zone samples (Figure C.4). However, the Bonanza samples contain abundant insect-pollinated taxa (i.e., *Pulsatilla* and *Rosa*), which are very poorly represented in the fossil samples. As a result, the apparent dissimilarity between the herb zone samples and Bonanza Bluff may only reflect the differing pollen collection and preservation characteristics of the sites, and not actual differences in the vegetation.

Additional analyses are needed before this question can be adequately answered. Several years worth of modern pollen should be collected from a variety of south-facing bluffs to smooth out differences due to yearly fluctuations of pollen production as well as differences due to variations in vegetation communities. In addition, the modern spectra should be compared with a number of lakes across Alaska and the Yukon Territory. The current analysis only studies two lakes, the apparent similarity of the late herb zone samples to Bonanza Bluff may be because all three sites are located in central Alaska. It is probable, because of regional differences in climate, that herb zone pollen from the Yukon, as well as the western Brooks Range is significantly different from Bonanza Bluff.

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